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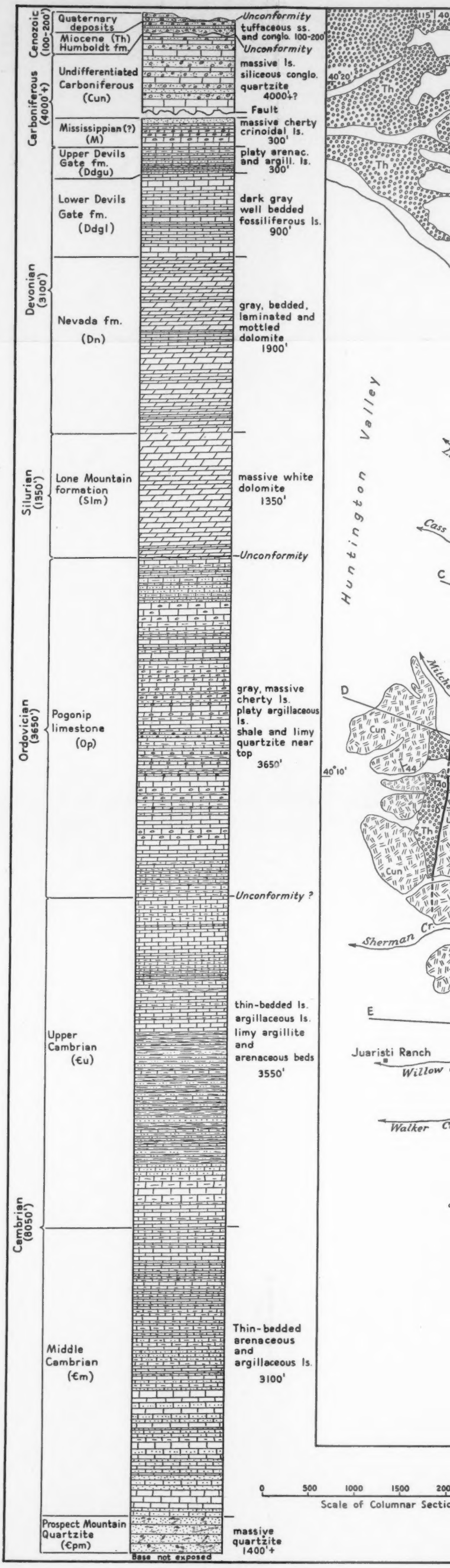
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Quaternary deposits are shown on Figure 2.

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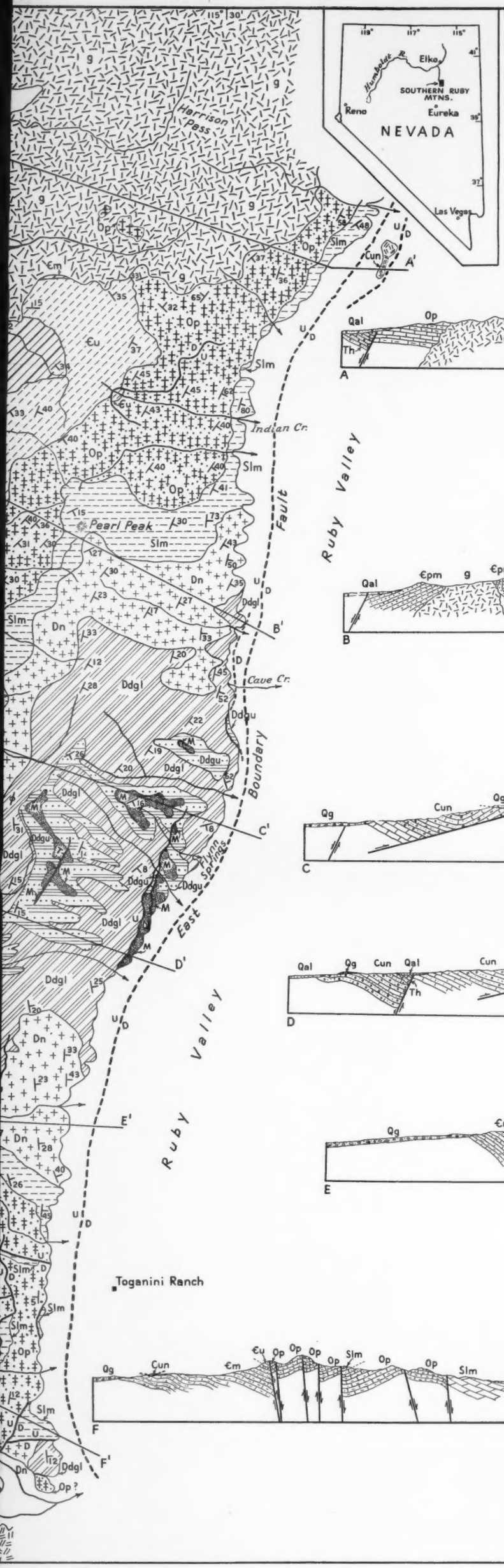
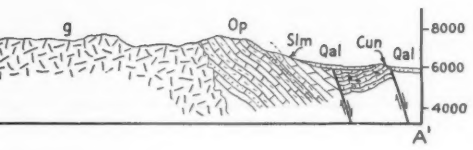


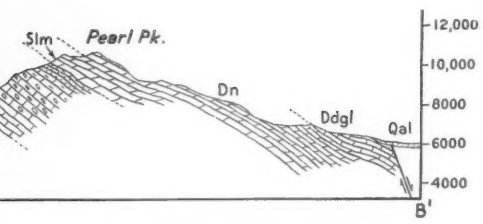
AND COLUMNAR SECTION OF SOUTHERN RUBY MOUNTAINS

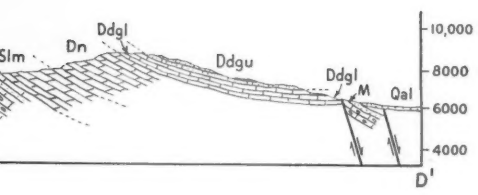
aternary deposits are shown on Figure 2.

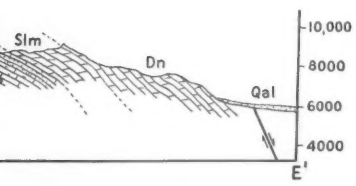
SOUTHERN
MOUNTAINS
NEVADA

ogy by R. P. Sharp, 1939









Th	Humboldt formation
Cun	Undifferentiated Carboniferous
M	Mississippian ?
Ddgu	Upper Devils Gate fm.
Ddgl	Lower Devils Gate fm.
Dn	Nevada formation
Slm	Lone Mountain formation
Op	Pogonip limestone
€u	Upper Cambrian fm.
€m	Middle Cambrian fm.
€pm	Prospect Mountain quartzite
g	granite intrusives

U / D — Normal Fault

Concealed Fault

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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STRATIGRAPHY AND STRUCTURE OF THE SOUTHERN RUBY MOUNTAINS, NEVADA

BY ROBERT P. SHARP

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ABSTRACT

A nearly complete Paleozoic section of exceptionally simple structure containing 16,450 feet of Lower Cambrian to Mississippian(?) beds is well exposed in the southern Ruby Mountains. In addition several thousand feet of undifferentiated Carboniferous rocks, partly Pennsylvanian, appear only in thrust sheets and fault blocks. The thickness and details of this section fit well with previously developed concepts of paleogeography in the northeastern Great Basin.

The internal structure of the range is a simple eastward-dipping homocline complicated to the north by granitic intrusions and to the south by a mosaic of high-angle normal faults. Along the west-central flank is a low-angle westward-

dipping thrust with a probable minimum eastward displacement of 7 to 10 miles. The thrusting can be dated with certainty only as post-Pennsylvanian and pre-Miocene but by analogy with other areas is probably early Tertiary.

The boundary structure passes from a westward-tilted horst, typical of the northern two-thirds of the range, to a simple westward-tilted fault block in the southernmost part. Movements along the boundary faults have occurred at intervals from Miocene to Pleistocene.

Early Wisconsin (Iowan) and later Wisconsin glaciers occupied the higher parts of the range and indicate a Wisconsin orographic snow line at approximately 10,000 feet. This snow line rose southward with a gradient of 200 to 300 feet per 100 miles.

INTRODUCTION

INTRODUCTORY STATEMENT

Many of the described Paleozoic sections in the Great Basin are located near mining camps which are almost necessarily areas of complex geologic structure, alteration, and igneous intrusion. Unavoidable errors of interpretation caused by such complications make highly desirable the study of Paleozoic formations in areas of more simple structure. The southern Ruby Mountains contain such a section.

This article describes: (1) the boundary structure of the southern third of the Ruby-East Humboldt Range, a representative Basin-Range block, and (2) its internal structure and stratigraphy. Particular emphasis is placed on the second matter as it is a foundation upon which interpretations of the metamorphic and igneous complex to the north must be based. As the rocks in the southern third of the range are well exposed, relatively unaltered, and largely devoid of igneous intrusions, this is an ideal place to establish a stratigraphic column that can be carried northward into areas of high-grade metamorphism and complex structure.

Nine weeks during June, July, and August 1939 were spent in the field, following shorter studies in preceding years. The relative simplicity of structure made for rapid work with a fair degree of detail and accuracy. Stratigraphic sections were measured by pace-and-compass traverses, and the field map was an enlarged copy of the Jiggs quadrangle on a scale of 1 inch to 4000 feet.

The project was supported by a grant from the Penrose Bequest of The Geological Society of America, which aid is gratefully acknowledged. Dr. T. B. Nolan and Mr. David Griggs kindly spent valuable time pointing out pertinent features of the Paleozoic section at Eureka, Nevada. Drs. C. W. Merriam, C. E. Resser, G. A. Cooper, A. H. Sutton, J. R. Ball, Josiah Bridge, H. W. Scott, and Mr. Maurice Smith studied fossil collections from this region, and their invaluable aid is greatly appreciated. Edward A. Schmitz assisted in drafting illustrations, and D. R. Hanna and J. E. Carlson of the National Youth Administration

aided in the office and laboratory. Max Arnhold, Herbert Dill, and Mr. and Mrs. Reginald Coffin extended many services and courtesies in the field. The manuscript has been improved by the careful critical treatment of Professor H. P. Woodward.

PREVIOUS WORK

Although scouts, trappers, and explorers like Ogden, Bonneville, Walker, Kern, Beckwith, and Fremont passed through this region between 1830 and 1860, Simpson (1876, p. 64, 331) and Engelmann, in 1859, were probably the first to record geological observations of the Ruby-East Humboldt Range. The most comprehensive early investigation was that of the Fortieth Parallel Survey in 1867-1873 (King, 1877, p. 525-541; 1878, p. 12, 62-70, 193, 195, 203-205, 475-476, 739). Wheeler (1889, p. 249-250) succinctly discusses the general relations within this region, and Spurr (1901, p. 222; 1930, p. 59-61) mentions the southern part of the range. The most up-to-date résumé is given by Hill (1916, p. 54-67); some notes on tungsten deposits have been published by Hess and Larsen (1921, p. 304-307). Blackwelder (1931, p. 910-911; 1934, p. 218-219) describes evidences of glaciation in the northern two-thirds of the range, and the writer has discussed the glaciation, adjoining basin deposits, boundary structure, and geomorphology of that area (Sharp, 1938; 1939a; 1939b; 1940).

PHYSICAL SETTING

The southern Ruby Mountains are located in Elko and White Pine counties, northeastern Nevada, approximately 80 miles west of the Utah border and 120 miles south of the Idaho line. (*See* inset of Plate 1.) They comprise that part of the Ruby-East Humboldt Range which extends from Harrison Pass southward for 22 miles to Overland Pass. This section of the range is a narrow, linear fault block that is widest and highest in its north and center areas. This rugged and asymmetrical mountain block has an abrupt east face (Fig. 1) and a more gentle western slope. It is bounded on the east by Ruby Valley, an interior basin with marshy bottom land—so-called Ruby Lake—and on the west by Huntington Valley which is drained by Huntington Creek, tributary to the South Fork of Humboldt River. The average altitude of the basins is about 6000 feet. As parts of the range reach more than 10,500 feet, the average maximum relief is about 4500 feet.

The climate is semi-arid. Perennial streams are moderately abundant on the western slope in the northern half of this area, but Cave Creek and the stream flowing from Flynn Springs are the only permanent streams on the eastern side.

STRATIGRAPHY

GENERAL INTRODUCTION

The Paleozoic section in the southern Ruby Mountains contains an unbroken sequence from Lower Cambrian to Mississippian(?) with every intervening system represented in nearly continuous outcrops. In the

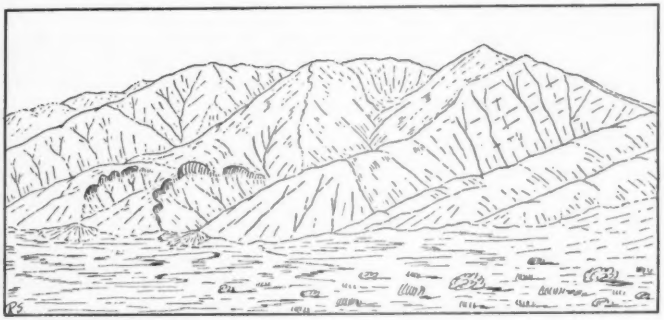

FIGURE 1.—*East face of southern Ruby Mountains*
Sketched from photograph near Pearl Peak, relief is about 4,800 feet.

central part these rocks are exposed as a simple eastward-dipping homocline, a fact that makes this place particularly attractive for stratigraphic studies. The total thickness of the continuous Paleozoic section is 16,450 feet, and to this can be added several thousand feet of undifferentiated Carboniferous beds separated from the normal sequence by faults. Igneous rocks and Cenozoic deposits are also present.

Insofar as reasonably possible, existing stratigraphic names have been applied to formations distinguished. Some formations are left unnamed to await possible future stratigraphic work and to avoid additions to the ever-lengthening list of stratigraphic names.

PROSPECT MOUNTAIN QUARTZITE (LOWER CAMBRIAN)

The Prospect Mountain quartzite was named by Hague (1883, p. 253-254; 1892, p. 34-35) from Prospect Peak in the Eureka district, Nevada. South of Harrison Pass this formation is exposed only on Rattlesnake Mountain dome at the mouth of Brown Creek (Pl. 1), where it consists of 1400 feet of brown massive quartzite. This exposure was incorrectly referred to the † Ogden quartzite by Hague (King, 1877, p. 529-531).

This formation is a massive vitreous quartzite (Pl. 2, fig. 1) that contains some schistose beds or partings and sparse conglomeratic layers. It is normally white, light gray, or light tan and weathers dark brown with associated streaks, spots, and

† Following U. S. Geological Survey practice a dagger is used to indicate abandoned terms.

diffusion rings of iron oxide. The quartzite is completely crystalline, is medium-grained (0.1 to 2 mm.), and contains minor amounts of orthoclase, zircon, apatite, ilmenite, muscovite, pyrite, and hematite or limonite, the first four of which appear to be detrital. Thin beds of mica schist scattered sparsely through the formation contain quartz, biotite, muscovite, sericite, sillimanite, zircon, and pyrite. This latter assemblage suggests high-grade metamorphism of shaly beds. The thin conglomerates contain small (½ to 1 inch) pebbles of whitish quartzite that have been secondarily flattened and elongated.

The Prospect Mountain is fairly well bedded on a 6-inch to 2-foot scale and near the top contains much thinner layers of mica schist. Some beds of quartzite are faintly laminated; others are crudely cross-bedded. The mica schist and parts of the quartzite show foliation parallel to bedding.

Calculations from structure sections and maps indicate a local thickness of 1400 feet, but greater thicknesses are reported elsewhere in Nevada and Utah (Lindgren and Loughlin, 1919, p. 22, 24; Nolan, 1935, p. 5; Wheeler and Lemmon, 1939, p. 17, 33; Wheeler, 1940, p. 16). The Prospect Mountain and related formations may be partly nonmarine as suggested by Walcott (1915, p. 183), Mansfield (1927, p. 180), and Deiss (1941, p. 1090).

No fossils were found in the formation in this area, where the oldest fossils collected (middle Cambrian) come from beds 1000 feet above the top of the quartzite. The Prospect Mountain is assigned to the Lower Cambrian on the basis of fossils found elsewhere near its top or in directly overlying strata (Butler, Loughlin, and Heikes, 1920, p. 79; Nolan, 1935, p. 6; Deiss, 1938, p. 1139, 1148-1149, 1157-1160; Wheeler and Lemmon, 1939, p. 17-18; Resser, 1939, p. 6). The basal Cambrian Brigham quartzite of northern Utah and southeastern Idaho may be wholly middle Cambrian (Deiss, 1938, p. 1118-1119, 1162; 1940, p. 783; Resser, 1939, p. 6) or, perhaps, partly early Cambrian (Blackwelder, 1910, p. 523; Butler, Loughlin, and Heikes, 1920, p. 78-79; Mansfield, 1927, p. 48; Eardley and Hatch, 1940, p. 836-837). As the early Cambrian marine invasion came from the south and southwest, the initial Cambrian rocks may be progressively younger toward the north. However, the quartzite in the southern Ruby Mountains should be largely early Cambrian because the entire Prospect Mountain quartzite at Eureka, 50 miles southwest, is early Cambrian (Wheeler and Lemmon, 1939, p. 17-18).

The Prospect Mountain may be correlated, in part, with the Tintic quartzite of Utah (Lindgren and Loughlin, 1919, p. 23; Butler, Loughlin, and Heikes, 1920, p. 78; Gilluly, 1932, p. 8, 18-19), with, perhaps, the Campito quartzite of eastern California (Knopf and Kirk, 1918, p. 28), and with the Stirling quartzite of southern Nevada (Nolan, 1929, p. 463) and southeastern California (Hazzard, 1937, p. 306). Its partial corre-

lation with the Brigham quartzite is controversial; Deiss (1938, p. 1161-1162) opposes it, but Eardley and Hatch (1940, p. 836-837) tacitly favor it.

MIDDLE CAMBRIAN FORMATION

Middle Cambrian beds are well exposed on various branches of Lindsay Creek, and a section along the ridge north of North Fork of Lindsay Creek is representative. The Middle Cambrian is also well exposed along the west base of the range south of Sherman Creek.

The Middle Cambrian consists of thin-bedded impure argillaceous and arenaceous limestone intercalated with limy shale and sandstone. Near igneous contacts many beds are altered to diopside granulite,[1] hornfels, quartz-biotite schist, argillite, marble, and quartzite. The total thickness is 3100 feet. Once, all beds in the southern Ruby Mountains above the Prospect Mountain quartzite were referred to the † Wasatch limestone, supposedly of Devonian and Carboniferous age (King, 1877, p. 528; 1878, p. 193, 203-204); now they are known to include Cambrian, Ordovician, Silurian, Devonian, and Mississippian(?) members.

In the north the basal beds are dark-gray coarsely crystalline massive limestone, with some lenses or beds of quartzite, succeeded by 400 to 500 feet of gray massive crystalline limestone with numerous arenaceous layers. The remaining part of the Middle Cambrian consists of impure arenaceous and argillaceous platy limestone variously colored white, gray, brown, pink, or green. Irregular argillaceous partings and micaceous layers are abundant, and the amount of sand is less than in the underlying beds. The uppermost 250 to 300 feet is light-colored relatively massive crystalline limestone with sparse argillaceous partings. It forms a prominent unit that is useful in field mapping.

Where altered by contact with igneous intrusions the Middle Cambrian contains diopside granulite, hornfels, quartz-mica schist, argillite, marble, and quartzite. Its many distinctive granulite beds are grayish green to nearly black, fine-grained, crystalline, and contain calcite and diopside with smaller amounts of quartz, biotite, and clinozoisite. Some of the large diopside porphyroblasts inclose quartz and calcite in a poikiloblastic arrangement. Locally near the top are granulites consisting of calcite with smaller amounts of quartz, biotite, hornblende, and traces of muscovite, plagioclase, and diopside. The presence of diopside indicates considerable magnesium carbonate in the original rock, although true dolomite may not have been present. Thin layers or lenses of biotite and quartz are associated with the granulites, and thicker beds of dark-gray to black quartzitic argillite are abundant near the base. Some argillite contains small lenticular aggregates of diopside, and many granulites are speckled with similar aggregates of biotite and diopside. The schistosity appears to be essentially parallel to the bedding.

The Middle Cambrian is unmetamorphosed and more argillaceous south of Sherman Creek, where the most common rock type is a platy, gray to brown, dense argillaceous limestone with buff argillaceous partings and yellowish limy shale.

[1] Granulite is used in the sense given by Holmes (1928, p. 111): "A granulose metamorphic rock, composed of even-sized interlocking granular minerals."

The argillaceous strata which weather brown or buff are locally splotched with red. Massive light-colored coarsely crystalline limestone likewise marks the top of the formation in this part of the range. As the formation is largely thin bedded, platy or flaggy outcrops are characteristic.

The following representative section was measured by a pace-and-compass traverse along the crest of the ridge north of North Fork of Lindsay Creek, starting at 8200 feet and continuing to the Prospect Mountain quartzite at the lower end of this same ridge. (Lithologic types are recorded as if unmetamorphosed.)

Middle Cambrian formation (descending order)	Feet
19. Limestone: gray, banded with white, massive, medium- to coarse-grained	17
18. Limestone: gray, crystalline, platy, with thin (¼ to ⅛ inch) regular argillaceous partings that weather dark brown	8
17. Limestone: gray, crystalline, massive, weathering to light colors; contains some irregular argillaceous accumulations and platy argillaceous limestone that weathers brown and pinkish	250
16. Limestone: light-colored, coarsely crystalline, massive, with intercalated layers of gray and pinkish platy argillaceous limestone	160
15. Argillaceous limestone: gray, green, pink, and white banded, finely crystalline, platy, with regular argillaceous and micaceous partings, and some beds of massive light-gray crystalline limestone	225
14. Limestone: greenish, finely crystalline, platy beds separated by chocolate-brown thin (⅛ to ¼ inch) sandy argillaceous bands, also occasional gray massive limestone containing irregular argillaceous and sandy masses, weathered surface pitted and knotty in appearance	155
13. Limestone: gray, finely crystalline, platy, with argillaceous partings which weather brown and a few beds of gray crystalline massive limestone	360
12. Argillaceous limestone: gray, green, brown bands, dense, with some extremely argillaceous layers	170
11. Limestone: greenish-gray, finely crystalline to dense, platy, with some light-gray thin limy quartzite bands	75
10. Limestone: gray, crystalline, massive, with some argillaceous platy limestone beds and a few argillites	50
9. Arenaceous and argillaceous limestone: gray, banded with white, platy, some layers notably micaceous	120
8. Limestone: gray, crystalline, thin-bedded, impure, containing thin arenaceous and argillaceous layers	405
7. Arenaceous limestone: black, somewhat massive, weathers black to dark brown	55
6. Arenaceous limestone: dark-gray, banded with light gray and brown, platy, with thin layers of greenish-gray limy quartzite and quartzitic sandstone which weathers brown	100
5. Quartzite: dark-gray, limy	5
4. Limestone: dark gray to black, finely crystalline, platy, impure, with intercalated dark limy quartzitic beds	200
3. Argillaceous limestone: black, finely crystalline, with dark quartzitic beds, more platy than in 4	295
2. Quartzite: black to dark gray, fine-grained, well-bedded, also some interbedded black impure limestone beds	170

Middle Cambrian formation (descending order)	Feet
1. Limestone: light to dark gray, coarsely crystalline, massive, with some beds of black impure limestone and irregular quartzitic lenses, the whole group weathers light gray	280
Total	3100

The Middle Cambrian overlies the Prospect Mountain quartzite with sharp but conformable contact. No beds resembling the Ophir, Cabin, or Pioche shale were found above the early Cambrian quartzite, although they are so reported in western Utah and eastern Nevada (Lindgren and Loughlin, 1919, p. 25-26; Butler, Loughlin, and Heikes, 1920, p. 78; Gilluly, 1932, p. 9-10; Westgate and Knopf, 1932, p. 8-10; Nolan, 1935, p. 6-7; Deiss, 1938, p. 1139, 1154-1156; Wheeler and Lemmon, 1939, p. 17-18). The top of the Middle Cambrian, as here defined, is at the top of a light-colored massive limestone that is overlain by thin-bedded limestone, quartzite, and argillite. Although this contact is a convenient field horizon, it has not yet been confirmed by fossil evidence in beds above.

Two collections of trilobites from the southern part of this area[2] have been identified by Dr. C. E. Resser as Middle Cambrian at about the *Bathyuriscus* or *Ehmania* zones. No fossils were found in the lowermost beds, so it is not impossible that they may be Lower Cambrian. Since this formation is probably equivalent to parts of several different Middle Cambrian units no attempts at definite correlation are made. Lithologic similarity between argillaceous beds of the Middle Cambrian in this area and those of the upper Secret Canyon shale at Eureka may be noted.

UPPER CAMBRIAN FORMATION

Above the Middle Cambrian and below the Ordovician is a rock section, 3550 feet thick, that contains late Cambrian fossils. These rocks are well exposed on branches of Lindsay Creek and at the heads of Pearl and Brown creeks (Pl. 5, fig. 2); a section along the ridge south of Pearl Creek is representative. The formation contains thin-bedded limestone, argillaceous and arenaceous limestone, and limy argillite.

The lower 700 feet of the Upper Cambrian is gray finely crystalline thin-bedded limestone and argillaceous limestone interbedded with argillaceous and arenaceous limestone that weathers buff and brown. Above is 1550 feet of dark-brown to black, thin-bedded argillite with interbeds of gray-green quartzitic argillite and thin gray limestone (Pl. 2, fig. 2). The succeeding 650 feet consists of gray finely crystalline thin-bedded limestone containing irregular argillaceous accumulations that may be indicative of shallow-water deposition. These argillaceous masses weather less readily than the limestone matrix to give the rock a distinct knotty appearance.

[2] One collection was made in argillaceous limestone beds at a small saddle 1 mile northwest of Cherry Springs, elevation 7150. The other collection came from white limestone beds 1 mile southeast of Cherry Springs at 7350 feet.

The top 650 feet consists of gray fine-grained platy limestone with some intercalated argillaceous material.

The Upper Cambrian becomes more argillaceous to the south, so that on Walker Creek it includes considerable greenish-gray thin brittle shale, containing large (2 inch to 1 foot) ellipsoidal limy concretions, many with trilobite remains. Similar concretionary shales extend north to Sherman Creek, where they give way to argillaceous limestone and limy argillite. Interbedded with the shale are thin gray nodular limestone beds and lenses. Gray dense nodular limestone overlies the shale.

Near intrusive contacts the Upper Cambrian beds have been altered to granulite similar to that of the Middle Cambrian and containing calcite, phlogopite, biotite, diopside, clinozoisite, tremolite, and oligoclase. Diopside grains may be clustered around large phlogopite crystals, and some of the limy argillites also contain aggregates of diopside. Farther from the intrusives the limestones are merely recrystallized and contain little diopside and phlogopite.

The Upper Cambrian is thin-bedded throughout, and color lamination or streaking is common in both limestone and argillite. Locally some of the argillite has a slaty cleavage.

The following representative section was measured by pace-and-compass traverse along the crest of the ridge between Brown and Pearl creeks starting at 9400 feet and extending down to the top of the underlying Middle Cambrian at 8500 feet.

Upper Cambrian formation (descending order)	Feet
10. Limestone: gray, finely crystalline, platy, with many moderately regular interbeds of argillaceous limestone which weathers brown or buff and is spotted with limonite	350
9. Limestone: gray, finely crystalline, thin-bedded (1 to 2 inches), with a few ½-inch argillaceous beds which weather buff. This group is less argillaceous than 10	140
8. Limestone: gray, black lines, finely crystalline, thin-bedded, with a few argillaceous beds	165
7. Limestone: light-gray, crystalline, thin-bedded, with numerous irregular argillaceous accumulations which weather brown and stand out on weathered surfaces giving a knotty appearance	170
6. Argillite: brown, fine-grained, impure, thin-bedded, with a weathered surface characterized by spherical pits	75
5. Limestone: gray, finely crystalline, thin-bedded, with irregular argillaceous masses which weather brown and stand out on the weathered surfaces giving a knotty appearance, also intercalated beds of impure fine argillite with spherical weathering pits	405
4. Argillite: dark-brown, nearly black, fine-grained, thin-bedded, also many grayish-green banded argillites and quartzitic argillites, also gray quartzite and occasional gray limestone beds with dark argillite displaying spherical weathering pits	1550
3. Limestone: gray, finely crystalline, thin-bedded, with interbedded 2- to 3-inch layers of grayish, brown-weathering, limy quartzite	135
2. Limestone: gray, finely crystalline, very thin-bedded, with intercalated beds of argillaceous limestone which weather brown, beds thicker toward the base	355
1. Limestone: light- to medium-gray, crystalline, thin-bedded, banded with somewhat irregular thin argillaceous layers which weather buff, some 2- to 3-inch argillite layers	205
Total	3550

The thickness of 3550 feet is representative in this area except toward the south where it is approximately 1600 feet. Lateral thinning and lensing of beds or an unconformity at the top of the Cambrian may account for this decrease. An unconformity between the Cambrian and Ordovician has been reported from western Utah (Richardson, 1913, p. 408; Lindgren and Loughlin, 1919, p. 29; Gilluly, 1932, p. 18-20; Nolan, 1935, p. 14-15) and southern Idaho (Mansfield, 1927, p. 173), but direct evidence of such an unconformity has not been found in this area.

Resser dates fossils from the middle of this formation as middle late Cambrian, Franconia division. Correlations with other Upper Cambrian formations in the Great Basin are not attempted.

POGONIP LIMESTONE (ORDOVICIAN)

The name Pogonip limestone was formerly used (King, 1878, p. 188) for all Cambrian and Ordovician beds between the Prospect Mountain and Eureka quartzites on Pogonip Ridge at White Pine, Nevada (Fig. 3). Hague (1883, p. 253-263; 1892, p. 48-54) adopted the name for Ordovician beds at Eureka, and it has since been applied to various Ordovician strata in the Great Basin (Spurr, 1903, p. 21; Ball, 1907, p. 28-29; Spencer, 1917, p. 24; Knopf and Kirk, 1918, p. 10, 34-35; Kirk, 1933, p. 32; Noble, 1934, p. 175; Hazzard, 1937, p. 322; Merriam, 1940, p. 8). Hague's Pogonip is largely Beekmantown, but beds near the top contain a Chazy fauna. Kirk (Knopf and Kirk, 1918, p. 34) suggested that Pogonip be restricted to the Beekmantown beds, but Walcott (1923, p. 466-467) preferred Pogonip for the upper part only. Holliday's (1940, p. 1984) report of middle or late Ordovician forms from the Pogonip at the type locality adds further complications. Since the U. S. Geological Survey still uses Pogonip in the broad sense (Wilmarth, 1938, p. 1690), application of this name to the Ordovician of the southern Ruby Mountains appears reasonable for the present. According to Wheeler and Lemmon (1939, p. 27-28) the lower 425 feet of the Pogonip limestone at Eureka is Upper Cambrian, but no Cambrian beds are known to be included in the formation here.

The Pogonip limestone is widely distributed on both flanks of the southern Ruby Mountains. Exceptionally well exposed sections can be seen at the head of Indian Creek (Pl. 5, fig. 1) and in Willow Creek; the latter is more representative. The Pogonip consists of 3650 feet of gray massive cherty limestone, thin-bedded argillaceous limestone, limy shale, intraformational limestone conglomerate, and near the top a distinctive limy sandstone and quartzitic member. The formation contains early and middle(?)Ordovician fossils.

The basal Pogonip is gray massive crystalline limestone with a few platy argillaceous beds that become more numerous southward. Considerable gray massive dense limestone containing layers, lenses, and nodules of black to gray chert constitutes the middle of the formation. These cherty beds are best exposed around Willow Creek (Pl. 2, fig. 3). To the north chert decreases and eventually disappears. The cherty limestone or equivalent beds are overlain by light-gray less massive limestone containing layers of intraformational limestone conglomerate distributed through several hundred feet. The conglomerate consists of elongated or tabular, subangular to rounded limestone pebbles, 1 to 2 inches in maximum dimension, set in a coarsely crystalline calcite matrix. No pebbles foreign to the immediate vicinity were recognized in the conglomerate, so it is probably intraformational and, perhaps, formed contemporaneously with similar conglomerates in other Ordovician formations of the eastern Great Basin (Richardson, 1913, p. 408; Lindgren and Loughlin, 1919, p. 31-33; Mansfield, 1927, p. 57, 181; Westgate and Knopf, 1932, p. 14; Nolan, 1935, p. 15).

The overlying beds are argillaceous limestone and greenish shale that pass northward into impure gray limestone. Near the top of the Pogonip is 250 to 300 feet of fine-grained limy sandstone and quartzite that constitutes the most distinctive marker band in the range. These strata contain about equal proportions of calcite and quartz with a few grains of detrital microcline and plagioclase. They are gray but weather brown with punky and porous surfaces owing to leaching. Prominent blocky talus slopes indicate outcrops of these beds, and locally they contain gastropods and small orthoid brachiopods. Arenaceous and argillaceous limestone is interbedded with the quartzite, and the remaining 250 feet above is gray, brown-weathering, thin-bedded arenaceous and argillaceous limestone. On the north fork of Mitchell Creek and just north of Overland Pass small outcrops of white massive vitreous quartzite at the top of the Pogonip resemble the upper Eureka quartzite.

Southward the Pogonip is locally a coarse-grained massive cherty brownish dolomite which weathers dark brown and contains irregular wormlike intergrowths of white coarse dolomite. A few oölitic limestone beds in the Pogonip do not appear to be extensive. Near intrusives the Pogonip shows effects of contact metamorphism including marmoritization and development of skarn minerals.

The following representative section was measured by pace-and-compass traverse along the ridge south of North Fork of Lindsay Creek from 9700 feet to the Upper Cambrian at 8100 feet.

Pogonip limestone (descending order)	Feet
19. Limestone: gray, dense, in beds a fraction of an inch to a few inches thick, with thinner more argillaceous layers which weather buff to brown, fossiliferous	225
18. Limy quartzite and sandstone: gray, weathers brown and usually porous and punky owing to leaching, intercalated with beds of gray dense platy limestone, argillaceous limestone, and sandy limestone, fossiliferous. Quartzite forms prominent talus slopes	260
17. Limestone: light- to dark-gray, massive, finely crystalline, containing irregular white calcite veinlets and locally chert, argillaceous and platy	265
16. Argillaceous limestone: gray, weathers brown, platy, and contains thick layers of chert locally	75
15. Limestone: gray, thin-bedded, with argillaceous partings	55

Pogonip limestone (descending order)	Feet
14. Limestone: dark-gray, thin-bedded, with irregular argillaceous partings, weathers buff	145
13. Limestone: gray, massive, with thin-bedded argillaceous limestone	170
12. Limestone: gray, massive, with some light-colored chert nodules, lenses, and layers	145
11. Limestone: gray, massive, dense to crystalline, containing considerable chert and some thin-bedded argillaceous limestone and minor amounts of argillite	510
10. Limestone conglomerate: gray, probably intraformational, intercalated in lined argillaceous limestone and gray and white massive limestone	65
9. Limestone: gray to light-colored, massive, coarsely crystalline, with light-gray cherty bands and some intercalated gray lined argillaceous limestone	260
8. Argillaceous limestone: gray, weathers brown, with much argillite and some massive coarsely crystalline cherty limestone	155
7. Limestone: light-colored, massive, crystalline, containing gray to brown chert	20
6. Limestone: gray, massive, crystalline, weathers light colored	130
5. Limestone: gray to whitish, massive, crystalline, containing some gray chert and some argillaceous layers	240
4. Limestone: dark-gray, pure, finely crystalline, homogeneous, in beds of moderate thickness	80
3. Limestone: gray, massive, medium-grained, crystalline, cut by white calcite stringers and veinlets	145
2. Limestone: gray, massive, and gray cherty limestone	115
1. Limestone: gray, massive, crystalline, brittle and locally platy, also gray massive limestone with thin interbeds of argillaceous limestone which weather buff and brown	590
Total	3650

The basal Pogonip is lithologically similar to the Upper Cambrian, and this contact is not sharp. However, the distinct difference between the Pogonip and the overlying massive Silurian Lone Mountain dolomite allows them to be separated with ease and accuracy. The absence of the Middle Ordovician Eureka quartzite from this area and the fact that platy limestone beds at the top of the Pogonip are missing in the south indicate an unconformity at the Ordovician-Silurian contact. The possibility of a similar unconformity is also recognized in near-by areas (Hague, 1883, p. 267; 1892, p. 57; Richardson, 1913, p. 408; Mansfield, 1927, p. 173; Kirk, 1933, p. 27, 39; Nolan, 1935, p. 17). The intraformational conglomerate within the Pogonip does not seem to indicate another unconformity but may mark changing conditions which led to deposition of the overlying argillaceous and arenaceous beds.

The Pogonip is probably thinner to the south than the 3650 feet measured on Lindsay Creek, but structural complications make this uncertain.

Trilobites collected near the base of the Pogonip were submitted to Resser, who identifies them as of the *Kainella* fauna (Canadian) rep-

resenting the lower Garden City or the Mons formation of Idaho and Utah. Cooper and Bridge identify trilobites, gastropods, and brachiopods from various parts of the formation as ranging from the *Nanorthis-Leiostegium* zone at the Cambrian-Ordovician boundary to the highest Pogonip, possibly Chazy, containing *Pliomerops canadensis* Billings, *Endoceras,* and *Leperditia.* The upper Pogonip in this area is definitely older than the Black River division of the Eureka quartzite.

Richardson (1913, p. 408-409) reports the Garden City formation of Utah as probably equivalent to part of the Pogonip, and other Ordovician units in Utah, Idaho, and Nevada may also be represented in this formation (Lindgren and Loughlin, 1919, p. 32; Nolan, 1935, p. 15).

LONE MOUNTAIN FORMATION (SILURIAN)

The Lone Mountain limestone was named by Hague (1883, p. 257, 262, 267; 1892, p. 57, 61) from Lone Mountain, 18 miles west of Eureka, Nevada. Merriam (1940, p. 13) reports that Hague's Lone Mountain limestone contains both Ordovician (Mohawkian) and Silurian beds. (*See also* Wilmarth, 1938, p. 1210.) He redefines the formation, changing the name to Lone Mountain formation, and limits it to barren massive dolomite beds below the Devonian Nevada formation and above the Roberts Mountains formation, a new Silurian unit. The Silurian of the southern Ruby Mountains belongs chiefly in the Lone Mountain formation as so redefined. These beds are exposed on both flanks of the range and underlie some of the higher crestline peaks. Good exposures may be seen near Pearl Peak and along Mitchell Creek. The formation contains 1350 feet of massive crystalline light-colored dolomite with a few poorly preserved Silurian fossils.

At the base is 100 feet of dark-gray to nearly black, medium-grained to dense, flaggy dolomite that weathers light gray and contains irregular intergrowths of white coarsely crystalline dolomite and pentamerid brachiopods in the upper part. The strata just above consist of light-gray medium- to coarse-grained crystalline dolomite that is distinguished by corals, crinoid stems, and scattered gray chert nodules. The lower part of the Lone Mountain consists of light-gray, nearly white-weathering, massive crystalline dolomite. Pearl Peak is composed of these beds, and abrupt cliffs or slopes and narrow, steep-walled canyons are abundant wherever they are exposed.

In the south much of the Lone Mountain is brownish-gray coarsely crystalline massive dolomite that weathers dark brown and contains black chert. Near Overland Pass this dolomite includes a few thin white quartzitic beds consisting of medium-sized rounded quartz grains in a dolomite matrix. Many quartz grains are secondarily enlarged.

The Lone Mountain is separated sharply from the underlying Ordovician, and the possibility of an unconformity at this contact has been

FIGURE 1. LOWER CAMBRIAN PROSPECT MOUNTAIN QUARTZITE
Near top of Rattlesnake Mountain dome.

FIGURE 2. THIN LIMESTONE AND ARGILLITE BEDS OF UPPER CAMBRIAN FORMATION
Argillite stands out on weathered surface. Middle Fork of Lindsay Creek.

FIGURE 3. CHERTY LIMESTONE OF ORDOVICIAN POGONIP
About 20 feet of strata exposed. Willow Creek.

FIGURE 1. "MACKEREL" DOLOMITE NEAR BASE OF DEVONIAN NEVADA FORMATION

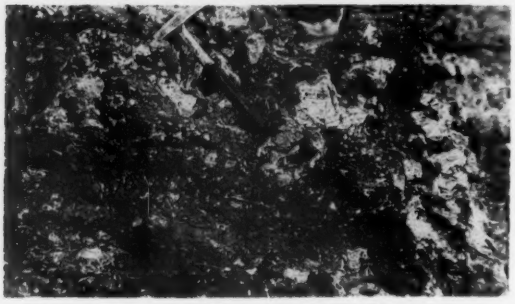

FIGURE 2. IRREGULAR CORAL-AND *Stringocephalus*-BEARING DOLOMITE
Near top of Devonian Nevada Formation.

FIGURE 3. ARGILLACEOUS LIMESTONE OF UPPER DEVILS GATE FORMATION

DEVONIAN FORMATIONS

discussed above. The contact with the overlying Devonian Nevada formation is not distinct. It has been placed within the change from light-colored massive dolomite to thinner flaggy beds of darker and finer-textured dolomite and limy dolomite of the Devonian.

Calculations from maps and structure sections give a mean thickness of 1350 feet. This is thicker than some Silurian sections in neighboring regions (Richardson, 1913, p. 407, 410; Mansfield, 1927, p. 58; Westgate and Knopf, 1932, p. 7; Nolan, 1935, p. 17) but not so thick as the Lone Mountain described by Merriam (1940, p. 19).

Ball identifies a few poorly preserved fossils from the Lone Mountain as probably early and middle Silurian. The fossiliferous strata have been found only in the lower 200 feet of the formation, and it is possible that they may belong in Merriam's (1940, p. 11-12) newly defined Roberts Mountains formation. Close lithologic similarity between the type Lone Mountain and the greater part of the Silurian in the southern Ruby Mountains and a similar relation to fossiliferous Silurian strata justify tentative correlation.

NEVADA FORMATION (MIDDLE DEVONIAN)

The term Nevada was first used by King (Atlas, map 4 of northeastern Nevada, 1876) for Chemung and upper Helderberg strata in the eastern Great Basin. The name was continued as Nevada limestone by Hague (1883, p. 264-266; 1892, p. 63) and as Nevada formation by Merriam (1936, p. 92), who redefines it to consist of Lower and Middle Devonian beds from the top of the Silurian to and including the *Stringocephalus* zone. Only middle Devonian fossils have been collected from the 1900 feet of strata mapped as Nevada in this area. The formation is exposed along the crest of the southern Ruby Mountains and on the east flank. Good sections can be seen north of Cave Creek and along the range crest south of Pearl Peak. It consists of various shades of gray crystalline, fine- to medium-grained, uniform to laminated or mottled, flaggy dolomite and limy dolomite. Particularly distinctive are beds of irregularly mixed black and white dolomite (Pl. 3, fig. 2) and *Stringocephalus*-bearing strata.

The Nevada formation is less uniform than the underlying Silurian. Colors range from nearly white to nearly black, and textures are dense to coarsely crystalline with a fine to medium grain predominant. Flaggy beds of dolomite and limy dolomite in various combinations of the above textures and colors constitute the major part of the formation. The lower Nevada on the range crest half a mile south of Pearl Peak contains 150 feet of gray to nearly black flaggy limestone with interbedded layers of buff- to red-weathering platy argillaceous limestone. This is one of the few localities where fossils are plentiful. A similar but thinner group of richly fossiliferous limestone beds appears in the formation farther south.

Distributed through the lower part of the formation are beds of nearly black crystalline "mackerel" dolomite, containing irregular, curving, oblong bodies of white crystalline dolomite some of which have fine-grained cream-colored cores of limy dolomite (Pl. 3, fig. 1). The white bodies are usually ¼ to ½ inch thick and 1 to several inches long, and some are thin bands that pinch and swell irregularly. Toward the top of the Nevada are massive coral-bearing beds composed of variously shaped masses of black and white crystalline dolomite in great confusion (Pl. 3, fig. 2). These distinctive beds are the best marker in the formation. Associated with the irregularly composed beds is *Stringocephalus*, whose value as an index fossil in Great Basin Devonian has been emphasized by Kirk (1927, p. 217-221) and Merriam (1940, p. 14, 16, 25).

Variations from the usual Nevada lithology in the south consist of massive gray-brown, dark brown-weathering, dense to crystalline dolomite and light-gray massive coarsely crystalline dolomite. In the north the uppermost Nevada is distinguished by 50 to 100 feet of brittle massive to platy and flaggy, white-weathering, limy dolomite interbedded with the usual gray crystalline dolomite. Thin groups of platy argillaceous beds near the base of the formation are too localized to be useful in mapping.

The following representative section was measured by pace-and-compass traverse along the crest of the range south from Pearl Peak.

Nevada formation (descending order)	Feet
27. Dolomite: gray, dense to crystalline, medium grained, with intercalated black limy dolomite	70
26. Limy dolomite: gray, dense, brittle, flaggy to platy	7
25. Dolomite: dark, crystalline, lined or streaked with irregular intergrowths of light coarsely crystalline dolomite	33
24. Limy dolomite: light-gray, dense, platy, with sparse intercalated beds of light coarsely crystalline dolomite and some dark dolomite	60
23. Dolomite: irregularly mixed black and white, massive, with corals	45
22. Dolomite: dark to medium-gray, irregularly composed, with some intercalated uniform dolomite beds	105
21. Limy dolomite: light-gray, dense, uniform	25
20. Limy dolomite: medium- to light-gray, dense to medium-grained, uniform, massive to flaggy, somewhat fossiliferous. Near top is some medium-gray irregularly composed dolomite	115
19. Dolomite: light- to medium-gray, dense to medium-grained, rather uniform, flaggy, some lined or laminated. At top is some light-gray coarsely crystalline dolomite which weathers light brownish	35
18. Limy dolomite: medium- to light-gray, dense, rather platy, laminated	50
17. Dolomite: light- to medium-gray, crystalline, medium- to fine-grained, uniformly laminated and somewhat streaked, flaggy, irregularly fractured, and a few scattered beds of dense limy dolomite, with faint mottling toward the base	115
16. Dolomite: dark, banded or streaked with light, also irregular dolomite intergrowths in bands alternating with lighter-colored more dense dolomite. Near middle is a dark and light irregularly composed dolomite bed containing *Stringocephalus*. A few scattered dark laminated dolomite beds included	135
15. Limy dolomite: brownish-gray, massive, irregularly mottled	10
14. Limestone: dark-gray, dense, with nearly black flaggy limestone containing many red- to buff-weathering platy argillaceous limestone beds, all more or less fossiliferous	165

Nevada formation (descending order)

	Feet
13. Dolomite: dark- to medium-gray, flaggy, somewhat mottled, grading down into dark-gray crystalline faintly streaked dolomite containing irregular white markings which may be fossils	125
12. Dolomite: dark, evenly streaked, also light-gray, white-weathering, dense dolomite	60
11. Dolomite: light-gray, crystalline, fine-grained, massive, laminated	5
10. Dolomite: dark, banded with faint irregular light-colored intergrowths, locally streaked, and some beds of light-colored dense limy dolomite, all irregularly fractured	60
9. Dolomite: light- to medium-gray, mottled, with several bands, 2 feet thick, of irregularly mixed light and dark crystalline dolomite (Pl. 3, fig. 1) and a few beds of white to light-gray dense faintly laminated limy dolomite	80
8. Dolomite: medium- to dark-gray, crystalline, laminated, irregularly fractured, also some light-, medium-, and dark-gray laminated flaggy dolomite, and some slightly mottled dolomite, all in beds 2 to 5 feet thick	65
7. Dolomite: light-, medium-, and dark-gray, crystalline, in bands 2 to 5 feet thick, with occasional beds of irregularly mixed light and dark crystalline dolomite	60
6. Dolomite: chiefly white to light-gray, massive, crystalline, coarse-grained, resembling the Silurian Lone Mountain, with occasional grayish irregularly composed thin (6-inch) dolomite beds, more dense toward the top	215
5. Limy dolomite: dark, mottled	15
4. Dolomite: light, somewhat mottled	90
3. Limy dolomite: gray, crystalline, medium-grained, flaggy, with an irregular texture and small vugs of quartz, also light-gray, almost white, faintly mottled flaggy dolomite with irregular areas of more coarsely crystalline dolomite, and a few beds of dark mottled limy dolomite	80
2. Dolomite: light-gray, white-weathering, dense, and dark dolomite in alternating bands 2 to 5 feet thick, also some beds of irregularly mixed light and dark crystalline dolomite	25
1. Limy dolomite: medium- to dark-gray, laminated and somewhat mottled, crystalline, fine-grained, with some lighter-colored limy dolomite and a few 3-foot beds of mottled irregularly mixed white and dark crystalline dolomite. Some beds in this group contain irregular vuggy growths of quartz and irregular lines of coarsely crystalline dolomite	50
Total	1900

The base of the Nevada is placed within the change from massive uniform light-colored Silurian dolomite to darker flaggy irregular dolomite and limy dolomite. The top is placed at the contact between dolomite and the limestone of the overlying Devonian Devils Gate formation. In places this contact is 500 feet above *Stringocephalus*-bearing beds. However, in this area *Stringocephalus* is distributed through several hundred feet, and according to Nolan (1935, p. 20-21) it may extend through a considerable thickness at Gold Hill, Utah. Therefore, choice of the upper limit of dolomite as the top of the Nevada does not involve

any serious departure from Merriam's definition of the formation (1940, p. 14). Field exposures and map relations indicate that the Nevada is conformable with the overlying Devils Gate formation. There are reasons for thinking that the Silurian and Devonian are separated by a disconformity or hiatus, but no proof was found here.

Fossils are sparse in the Nevada except in limestone beds. Merriam identifies the following forms from the limestone and reports that they correspond to his *Martinia kirki* zone of the type upper Nevada: *Martinia kirki* Merriam, *Atrypa* sp. b (large), *Leiorhynchus* sp. a, *Leiorhynchus* cf. *L. castanea* (Meek), and what appears to be a new species of the trilobite genus *Proetus*. *Stringocephalus* and *Favosites* cf. *F. limitaris* Rominger have also been found near the top of the formation. Only middle Devonian fossils have been found in the Nevada of this area, although elsewhere it also contains early Devonian forms (Merriam, 1940, p. 8). Lack of fossils from lower faunal zones may be due to the extensive dolomitization, or the lower beds may be missing, for the 3100 feet of Devonian strata here is 1400 feet less than the total reported by Merriam (1940, p. 14) from the Roberts Mountains region.

DEVILS GATE FORMATION (MIDDLE AND UPPER DEVONIAN)

General statement.—The Devils Gate formation has been named by Merriam (1940, p. 16) from Devils Gate 8 miles northwest of Eureka, Nevada. This formation is represented in the southern Ruby Mountains by 1200 feet of Middle and Upper Devonian strata. Fossils of the *Spirifer argentarius* and *Cyrtospirifer* zones and probably the *Phillipsastraea* zone (Merriam, 1940, p. 8) of the type Devils Gate are present. The upper 300 feet is easily separated on lithology, so two divisions have been mapped and described separately. The lower division is Middle Devonian, and the upper division is Upper Devonian.

Lower division of Devils Gate formation.—The lower Devils Gate covers a large area in the southern Ruby Mountains and is well exposed on the east face south of Cave Creek (Pl. 4, fig. 1). The formation consists of dark-gray to black, dense to crystalline, flaggy to massive limestone beds that contain abundant fossils including widely distributed *Cladopora* and stromatoporoids.

The lower Devils Gate is composed of rather pure uniform limestone except for a few thin beds of dense light-colored limy dolomite at the base. Colors are dark gray to black. Weathered surfaces are lighter and are perforated by a multitude of small cylindrical openings that have been developed by leaching during weathering. Textures are crystalline in fine to medium sizes with a few dense and even fewer coarsely crystalline beds. Some limestone in the middle of this division is oölitic, and many beds toward the top are laminated. The lower Devils Gate is well strati-

fied in beds a few inches to 2 feet thick (Pl. 4, fig. 1), and the few more massive cliff-forming units are prominent.

A highly characteristic feature is the presence of stromatoporoids and *Cladopora* which are widely recognized in Great Basin Devonian (Knopf and Kirk, 1918, p. 37; Nolan, 1935, p. 20-21; Hazzard, 1937, p. 331; Stumm, 1937, p. 424; Merriam, 1940, p. 15, 16). These forms are limited to the lower Devils Gate in the southern Ruby Mountains but have been reported from the Nevada formation elsewhere (Stumm, 1937, p. 424; Merriam, 1940, p. 15). The lower Devils Gate also contains a variety of brachiopods, corals, and gastropods, most of which have been replaced by white calcite and, therefore, stand out prominently in the dark limestone.

The following representative section was measured by a pace-and-compass traverse in the south branch of the canyon southwest of Cave Creek.

Lower division of Devil's Gate formation (descending order)	Feet
14. Limestone: dark-gray to black, finely crystalline, flaggy, fossiliferous	30
13. Limestone: black, dense, brittle, flaggy, with irregular fracture	35
12. Limestone: gray, massive, crystalline, with irregular fracture, cliff-forming unit	60
11. Limestone: dark- to medium-gray, crystalline, massive, weathers light gray and irregularly pitted, cliff-forming unit	50
10. Limestone: medium- to light-gray, massive, crystalline, fine- to medium-grained, many beds laminated and contain small stromatoporoids	25
9. Limestone: gray, massive, of irregular texture and irregularly weathered surface, contains large stromatoporoids and large gastropods. Massive beds are separated by flaggy beds of laminated crystalline limestone, whole cliff-forming	80
8. Limestone: brownish-gray, dense, massive to flaggy, with sparse stromatoporoids, some brachiopods, and many *Cladopora*, includes some black dense flaggy limestone which weathers irregularly and is somewhat streaked, also occasional massive beds of crystalline limestone	60
7. Limestone: dark-gray, crystalline, medium-grained, containing *Cladopora* and interbedded with black dense flaggy limestone with light-weathering intergrowths, some beds of gray crystalline limestone containing stromatoporoids and some beds with sparse brachiopod shells	110
6. Limestone: black and dark- and light-gray, dense, massive to flaggy, with some stromatoporoid and *Cladopora*, a few brachiopod remains and locally irregularly mixed light- and dark-gray limestone	75
5. Limestone: dark-gray, dense to finely crystalline, flaggy, locally platy and lined, considerable *Cladopora* and small-scale irregular nodular structure	70
4. Limestone: light-gray, massive, and dark-gray, flaggy, with *Cladopora*, locally some black dense stromatoporoid and *Cladopora* limestone with irregular red and white fine-grained areas all cut by veinlets of white calcite	60
3. Limestone: black and gray, brittle, flaggy, locally some light-gray dense to coarsely crystalline massive limestone	60
2. Limestone: dark-gray, dense, with irregular lighter-colored intergrowths of crystalline calcite, some stromatoporoids and *Cladopora*	70
1. Limestone: dark- to light-gray, massive, flaggy and platy, with stromatoporoid and *Cladopora* limestone, also occasional beds of nearly white dolomitic limestone indicating transition to underlying Nevada formation	115
Total	900

The lower contact is determined by an abrupt but seemingly conformable change from dolomite to limestone. The upper boundary is placed within the change from relatively pure limestone to arenaceous and argillaceous limestone belonging to the upper division. The thickness of 900 feet is representative throughout the area of exposure.

Merriam (1940, p. 8) recognizes three fossil zones in the type Devils Gate of which only the lower or *Spirifer argentarius* zone is represented in the lower Devils Gate of this area. Collections from the lower part of this zone have been studied by Merriam, who identifies *Spirifer engelmanni* Meek, *?Platyschisma mcCoyi* Walcott, *Atrypa* sp., *Cladopora* sp., and *Martinia* cf. *M. nevadensis* (Walcott). A collection from the upper part of the zone contains *Spirifer utahensis* Meek and *Atrypa* cf. *A. montanensis* Kindle. The lower division of the Devils Gate is Middle Devonian and appears to correspond approximately with the Middle Devonian part of the type Devils Gate. For a discussion of correlation between this and other Devonian formations the reader is referred to Merriam (1940, p. 61-77).

Upper division of Devils Gate formation.—This formation is best exposed on the east face of the range south of Cave Creek and on the range crest at Mitchell Creek. It consists of platy arenaceous and argillaceous limestone which weathers a brown or red (Pl. 3, fig. 3) and contains lenses of chert-pebble conglomerate near the base.

This division is distinguished by an abundance of arenaceous and argillaceous limestone. Most beds are gray, and weathered surfaces are brown or red. Well-rounded, medium-sized quartz grains compose up to 40 per cent of some beds. Siliceous conglomerate lenses at the head of Mitchell Creek are several feet thick and consist of rounded to subangular ½- to 1½-inch pebbles of brown, gray, and white quartzite and gray, green, and nearly black chert. The upper Devils Gate is not fossiliferous except toward the base. The following representative section was measured by pace-and-compass traverse at the head of the south branch of the canyon southwest of Cave Creek.

Upper division of Devil's Gate formation (descending order)	Feet
8. Limestone: black, buff-weathering, platy, underlain by a group of limestone beds of similar character interbedded with gray arenaceous limestone layers, 6 inches to 1 foot thick, which weather buff and rest upon a group of black and gray paper-thin limestone and argillaceous beds, buff-weathering, with thin black bedded chert layer at base	100
7. Limestone: gray, platy, argillaceous, locally highly arenaceous, weathers brown	25
6. Arenaceous limestone: gray, platy, weathers brown to reddish	20
5. Arenaceous and argillaceous limestone: gray, platy	25
4. Limestone: gray, flaggy, and arenaceous limestone which weathers brown and punky, some argillaceous platy limestone	45

Upper division of Devil's Gate formation (descending order)	Feet
3. Limestone: gray, finely crystalline, flaggy, with some beds of gray arenaceous limestone which weathers brown to red and punky, some platy limy argillite beds	40
2. Limestone: black to gray, platy, and some platy limy argillites which weather punky and brown, sparse arenaceous beds	25
1. Limestone: dark-gray, very thin platy beds, with gray arenaceous limestone and limy argillite beds which weather red and brown	20
Total	300

Contact between the two divisions of the Devils Gate formation is transitional. Local lenses of chert pebble conglomerate near the base of the upper division do not contain fragments of the underlying beds and probably indicate changes only in source and conditions of deposition. The contact with the overlying Mississippian(?) shows no evidence of erosion, although an unconformity at the base of the Mississippian has been recognized in southern Idaho and western Utah (Lindgren and Loughlin, 1919, p. 36-38; Mansfield, 1927, p. 173, 182; Gilluly, 1932, p. 22; Nolan, 1935, p. 21-22).

Merriam identifies the following forms from the upper Devils Gate in this area: *Cyathophyllum* sp., *Syringopora* sp., *Productella* sp., *Schizophoria simpsoni*, and *Schuchertella* cf. *S. haguei*. These fossils belong to the *Cyrtospirifer* zone of the Upper Devonian part of the type Devils Gate and may possibly represent in part the Upper Devonian *Phillipsastraea* zone.

UNDIFFERENTIATED DEVONIAN

Two isolated masses of gray to brown, dense to coarsely crystalline limestone containing abundant corals are associated with Carboniferous strata in an overthrust sheet near Overland Pass. Merriam identifies corals from the easternmost mass as probably *Pachyphyllum* from the *Phillipsastraea* zone of the early Upper Devonian part of the Devils Gate. These beds differ so in fossil content and lithology from other Devonian strata in the range that they must have been introduced from some remote point by thrusting. For this reason they have not been included in the normal stratigraphic sequence and are merely shown as Devonian on the map (Pl. 1).

MISSISSIPPIAN(?) FORMATION

Massive dark-gray to black cherty crinoidal limestone above the upper Devils Gate has been mapped separately and is tentatively classed as Mississippian. The thickness of 300 feet includes some brown, black, and green shale and gray quartzite overlying the crinoidal limestone. This formation occupies the highest position in the normal stratigraphic sequence and is exposed only on the east face of the range in small isolated areas.

At the base is 45 feet of dark-gray nodular limestone that weathers mouse gray. Near the base the matrix between nodules is brownish-green and clay rich, but higher up it is dark limy material. Many of the nodules are fossiliferous. Most of the formation consists of dark-gray medium- to coarse-grained limestone banded with nodular layers and lenses of gray and black chert. The limestone layers are 2 to 6 inches thick, and locally nearly 50 per cent of the rock is chert. The uppermost limestone beds contain abundant crinoid stems and sparse fragments of other fossils. They are overlain by 30 to 40 feet of brown, black, and green noncalcareous shale and gray noncalcareous quartzite that weathers brown. This quartzite is locally coarse-grained and contains about 25 per cent small rounded granules of red, brown, and black chert. These clastic beds form only small residual patches above the limestone. The following representative section was measured by hand leveling on the north flank of the prominent flat-topped ridge (Pl. 4, fig. 2) northwest of Flynn Springs, except for unit 10 which is added from other exposures.

Mississippian (?) formation (descending order)	Feet
10. Shale: brown, black, and green, and gray quartzite which weathers brown and is locally gritty with small rounded granules of black and brown chert	30
9. Limestone: light-gray, crystalline, banded with black chert layers and nodules at 3- to 6-inch intervals; nearly 50 per cent of the rock is chert; abundant crinoid stems	27
8. Limestone: gray, dense, flaggy, noncherty	17
7. Limestone: gray, dense, with black nodular chert layers, ½ to 1 inch thick along bedding at 6-inch intervals, many crinoid stems and some brachiopod fragments	40
6. Limestone: gray, flaggy, no chert	14
5. Limestone: gray, crystalline, thick-bedded, massive, uniform	16
4. Limestone: light-gray, in 6-inch beds with 1- to 2-inch layers of black chert; near top chert is less pure and weathers buff and punky, also 2 to 3 feet of irregular sandy and cherty limestone at top	40
3. Limestone: dark-gray, crystalline, massive, uniform, and noncherty	11
2. Limestone: gray, in beds 6 inches to 1 foot thick, banded with irregular nodular layers of black chert, 1 to 2 inches thick, chert rather calcareous toward top and weathers brown and punky	60
1. Limestone: dark-gray, banded, nodular, brownish-green argillaceous material between nodules near base, nodules 1 to 3 inches in diameter and frequently fossiliferous; one massive 5-foot uniform bed near top	45
Total	300

A distinctive feature of this formation is its cliff-forming habit (Pl. 4, fig. 2). The thickness of 300 feet measured above is a maximum.

Fossils are abundant in the nodular limestone near the base but sparse in overlying beds except for crinoid stems. Merriam has not been able to recognize anything strictly Devonian among these fossils, and Sutton reports that a *Syringopora* collected 150 feet above the base has a Mississippian aspect. For these reasons the strata are tentatively assigned to the Mississippian.

FIGURE 1. EASTWARD-DIPPING LIMESTONE. LOWER DEVILS GATE FORMATION
View northward on range crest west of Cave Creek.

FIGURE 2. MASSIVE MISSISSIPPIAN (?) LIMESTONE
Gentle slopes beneath cliff underlain by upper division of Devonian Devils Gate formation. View eastward toward Ruby Valley near Flynn Springs.

FIGURE 3. GRANITE OF HARRISON PASS STOCK AT HEAD OF CORRAL CREEK

DEVONIAN, MISSISSIPPIAN (?), AND GRANITE

FIGURE 1. EASTWARD-DIPPING LIMESTONE BEDS OF ORDOVICIAN POGONIP
View southward along crest of range at head of Indian Creek.

FIGURE 2. GLACIAL CIRQUE AT HEAD OF BROWN CREEK
Rocks are Cambrian and Ordovician. Small glacial pond at lower right.
View eastward.

FIGURE 3. LATER WISCONSIN GLACIAL MORAINE, CANYON
EAST OF PEARL PEAK

Merriam (1940, p. 40) reports massive cliff-forming crinoidal limestone above the Devils Gate at Dutch John Mountain near Pioche and at Newark Mountain near Eureka (personal communication) which may be either Carboniferous or Devonian. The shale and quartzite above the crinoidal limestone in the southern Ruby Mountains are similar to beds in Merriam's (1940, p. 28) Lower Carboniferous Diamond Peak series.

UNDIFFERENTIATED CARBONIFEROUS STRATA

General statement.—Carboniferous rocks other than the Mississippian (?) just described are exposed in three areas: (1) along the west flank of the range between Cass House and Sherman creeks, (2) in Overland Pass, and (3) at the east base of the range south of Harrison Pass. In each place the rocks are exposed in thrust sheets or in fault blocks, so their position in the normal stratigraphic sequence is not known. Each area will be considered independently.

Cass House Creek-Sherman Creek area.—The Carboniferous here composes a large thrust sheet. Siliceous chert-pebble conglomerate, quartzite, and massive limestone are abundant in a thickness estimated between 4000 and 10,000 feet.

The conglomerate consists of subangular to rounded pebbles of brown, red, gray, black, white, and green chert, and brown, tan, and white quartzite. The ratio between chert and quartzite is about 3 to 1, and the average pebble size is ½ to 1 inch. Associated with the conglomerate are numerous beds of noncalcareous vitreous white, gray, and brown coarse-grained quartzite containing a few chert granules and small pebbles. Extremely massive beds of grayish-brown crystalline limestone, usually badly fractured and healed with veinlets of white calcite, are interbedded with the clastic deposits. These beds contain light-brown chert in irregular nodules and layers and are locally fossiliferous. Coarse oölitic limestone and dense black flaggy to platy argillaceous limestone are less abundant. Limestone is dominant in the northern and southern parts of this area of Carboniferous rocks, and conglomerate and quartzite compose about 70 per cent of the section near Mitchell Creek.

Lack of exposures makes accurate determination of the thickness impossible, but calculations based on outcrop breadth and uniform dip of 30° E. give a maximum of 11,400 feet. If the structure is complicated by undetected folds and faults, as assumed in sections CC′ and DD′ of Plate 1, the thickness is probably less. A conservative estimate would be between 4000 and 10,000 feet.

Sutton suggests that poorly preserved fossils from the limestone are definitely Carboniferous and probably pre-Permian. Fusulines in beds near the top have been dated by H. W. Scott as Pennsylvanian. The

Carboniferous beds in the Cass House-Sherman creeks area may be both Mississippian and Pennsylvanian or, perhaps, wholly Pennsylvanian.

Overland Pass area.—The Carboniferous beds in Overland Pass are part of the thrust sheet exposed in the Cass House-Sherman creeks area and likewise consist of interbedded conglomerate, quartzite, and limestone.

The conglomerate consists of chert and quartzite pebbles like those already described and forms dark-brown bold outcrops. The quartzite beds are gray to nearly white, fine- to coarse-grained, and contain subangular quartz grains with traces of muscovite and iron oxides. The limestone is massive, gray, and brownish gray, fine- to coarse-grained, and badly fractured and healed with white calcite veinlets. A few beds of nearly white pure limestone and dense, black to dark-gray, flaggy to platy, impure argillaceous limestone with scattered chert nodules and layers have been observed.

The total thickness exposed appears to be several thousand feet, but complex structure prevents accurate measurement. Close lithologic resemblance to the strata in the Cass House-Sherman creeks area suggests that these rocks are also Carboniferous.

Area at east base of range.—Limestone is exposed in a low, elongate knoll ½ mile east of the range base just south of Harrison Pass (Pl. 1). A fault separates these rocks from Silurian and Ordovician strata composing the main body of the mountains.

They consist of gray and light-brownish gray, flaggy to massive, fine- to medium-grained limestone locally argillaceous and containing considerable gray chert as nodules and irregular bedding lenses. Nearly black dense carbonaceous limestone, fractured and healed with veinlets of white calcite, is less abundant. Nearly all these strata weather a light gray or muddy tan.

Sutton suggests that fossils collected at this locality are certainly Carboniferous and have a definite Pennsylvanian aspect. Spurr (1903, p. 60) probably collected his Coal Measure fossils here.

HUMBOLDT FORMATION (MIOCENE)

The Miocene Humboldt formation[3] (Sharp, 1939a, p. 136) consists of terrestrial deposits of fluviatile and lacustrine origin containing considerable volcanic material. This formation is exposed in a small fault block at the west base of the range between Mitchell and Sherman creeks and in Huntington Valley north of Pearl Creek. At both places the rocks are fine tuffaceous conglomerate and coarse tuffaceous sandstone.

[3] The Humboldt formation has previously been dated as upper Miocene (Sharp, 1939a, p. 151-154), but R. W. Chaney advises that plant remains collected near the base may be as old as early Miocene.

The conglomerate contains angular fragments of quartz, biotite, and white, pink, and green rhyolitic tuff, and the sandstone contains quartz and tuffaceous materials. These deposits are water laid but show only crude stratification, some cross-bedding, and poor sorting. They are weakly cemented and weather brown from an original light gray.

Not more than 200 feet of the formation is exposed here, but elsewhere it is several thousand feet thick. The Humboldt rests unconformably on pre-Tertiary rocks.

QUATERNARY DEPOSITS

General statement.—Quaternary materials in the southern Ruby Mountains area consist of fanglomerate, drift, and beach, pediment, and terrace gravels in addition to Recent stream and slope-wash gravels and alluvium (Fig. 2).

Fanglomerate.—This deposit is exposed along the west flank of the range between Lindsay Creek and Overland Pass (Fig. 2) and extends westward into Huntington Valley. Spurr (1903, p. 61) mistook it for the Miocene Humboldt formation. The thickness increases from a feather edge well up on the mountain block to an observed maximum of 150 feet.

In some places the fanglomerate is composed solely of limestone fragments, in others it contains conglomerate, quartzite, and limestone, and elsewhere it consists of granite, quartz, granulite, and limestone. The fragments are angular, subangular, and subrounded, an inch to several feet in diameter, and firmly cemented in a sparse matrix by calcareous material. The color differs with the constituents, but some shade of gray prevails. Sorting is poor, and the extremely crude bedding is inclined gently valleyward.

This material was deposited along the flanks of the present range by ephemeral flood-water streams. It has subsequently been dissected on the mountain block and buried toward the center of Huntington Valley. Some streams have cut through the fanglomerate and 400 feet into bedrock leaving it as a capping on ridges. The fanglomerate is attributed to uplift of the mountain block by faulting presumably along the east boundary fault, because at the west base there is no evidence of fault movement immediately preceding deposition of the fanglomerate. This uplift accompanied by westward tilting would rejuvenate the streams in the mountains but not in the valley to the west so they would be forced to drop their newly acquired load along the west base of the range. The upper edges of the fans so constructed would encroach upon the mountains as the deposits accumulated. Dissection of the fanglomerate on the mountain block and its burial beneath the alluvium of Huntington Valley can be attributed to subsequent uplift with westward tilting. Absence of Quaternary fanglomerate to the

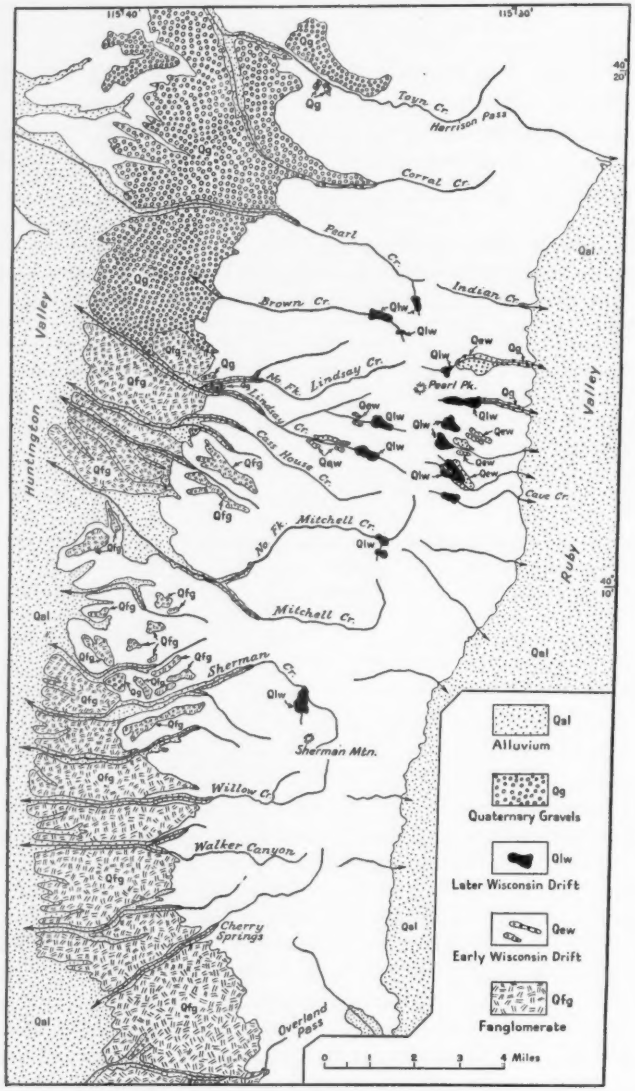

FIGURE 2.—*Map of Quaternary deposits*

north may be due to the ability of the permanent streams in that area to handle the increased load.

The age of the fanglomerate cannot be precisely determined. Depth of dissection shows that it is older than the Wisconsin glaciation in the range, and stratigraphic and structural relations make it younger than the Miocene Humboldt formation. The last major uplift of the Ruby-East Humboldt block occurred in late Pliocene to mid-Pleistocene time (Sharp, 1939b, p. 902-904). Since the fanglomerate may be related to any single uplift within this period, it may have been formed any time during late Pliocene and early Pleistocene.

Glacial drift and gravels.—Two glacial tills have been mapped in the southern Ruby Mountains (Fig. 2). These deposits have the usual bouldery character of mountain tills and form long lateral ridges on canyon walls and terminal or recessional loops across canyons (Pl. 5, fig. 3) and the mouths of cirques.

Outwash gravel has been mapped in canyons on the east side of the range (Pl. 5, fig. 3). Fan and pediment gravels skirt the base of the range on both sides and in places are transitional into terrace and modern stream gravels. Beach gravel was deposited along the shores of a Quaternary lake occupying Ruby Valley and is particularly abundant just south of Harrison Pass. This material consists of worn well-sorted tabular limestone pebbles.

IGNEOUS ROCKS

General statement.—Igneous rocks are common only in the north. Wholly intrusive, they are treated under the following heads: (1) Harrison Pass stock, (2) Rattlesnake Mountain dome intrusive, (3) dikes. Close geographic association and petrographic similarity suggest that these bodies are related to a single period of igneous activity.

Harrison Pass stock.—Harrison Pass is located at the center of a nearly circular area of granite 7 to 8 miles in diameter. Size, shape, and discordant intrusive contacts indicate that this body is a stock.

Gray coarse-grained to porphyritic biotite granite which locally contains phenocrysts of orthoclase 1 to 2 inches long is the most common rock type. It has approximately 20 per cent quartz, 45 per cent orthoclase, 33 per cent oligoclase, 2 per cent biotite, and traces of zircon and apatite. The plagioclase is well zoned. Irregularly distributed along the borders are masses of medium-grained muscovite-biotite granite suggesting that this may be a multiple stock. These two rock types are not separated on the geologic map (Pl. 1), and the order of their intrusion is not certainly known. Against limestone wall rock the granites may be green and contain considerable amounts of epidote, sphene, rutilated quartz, and a higher than average percentage of plagioclase. Assimilation of limestone by the magma is suggested.

Rattlesnake Mountain dome intrusive.—An igneous body, 1 mile long by ½ mile wide, is exposed in the core of the Rattlesnake Mountain dome on Brown Creek (Pl. 1). This intrusive consists of gray medium-grained biotite granite [4] and white medium-grained muscovite granite.[5] These two rock types have not been mapped separately, and there is conflicting evidence as to relative age. This body is discordantly intrusive into the Cambrian quartzite composing the dome, so it is possibly the top of a small multiple stock.

Dikes.—Numerous irregular dikelike bodies of pegmatite cut the Harrison Pass stock, and one pegmatite near the head of Corral Creek contains lepidolite. Several porphyry dikes, 5 to 20 feet thick and 1 to 2 miles long, have been mapped south of Harrison Pass. One particularly prominent dike in Lindsay Creek consists of biotite granite porphyry,[6] and another dike south of the east end of Harrison Pass is composed of quartz monzonite porphyry.[7] Smaller dikes of white rhyolite porphyry were also observed. High on the east face of the range north of Cave Creek are several dikes of augite-hornblende diabase.

Age of igneous rocks.—Except, possibly, for the diabase and rhyolite dikes, the various bodies are assigned to a single period of igneous activity, of which, however, they may represent different stages. Field evidence fails to date these bodies more precisely than post-Carboniferous and pre-Miocene. Certain analogies, however, may be made; Ferguson (1929, p. 117) suggests intrusions in southern and eastern Nevada are no older than Eocene. Many approximately early Tertiary intrusives are reported within 200 miles of the Ruby Mountains (Lindgren and Loughlin, 1919, p. 103-104; Butler, Loughlin, and Heikes, 1920, p. 99-100; Gilluly, 1932, p. 69; Westgate and Knopf, 1932, p. 20, Pl. 1; Nolan, 1935, p. 43, 48; Merriam, 1940, p. 30), and post-early Cretaceous intrusions are known at Eureka (Nolan, personal communication). Great Basin Tertiary intrusives are supposedly more granitic than those of Jura-Cretaceous age, which approximate a granodiorite and are deficient in pegmatite (Lindgren, 1915, p. 255). The granitic composition and location of the Ruby Mountains intrusives may be

[4] This rock contains approximately 15 per cent quartz, 65 per cent orthoclase and microcline, 13 per cent sodic oligoclase, 7 per cent biotite, and traces of magnetite and zircon. Potash feldspar is somewhat perthitic, plagioclase is zoned, and quartz is locally granulated.

[5] This rock contains approximately 30 per cent quartz, 60 per cent microcline, 5 per cent sodic oligoclase, 5 per cent muscovite, and traces of magnetite and garnet.

[6] This rock contains approximately 20 per cent quartz, 40 per cent orthoclase, 30 per cent oligoclase, 10 per cent biotite, and traces of muscovite, chlorite, and epidote. A myrmekitic, almost micropegmatitic, groundmass is particularly noticeable.

[7] This rock contains approximately 30 per cent quartz, 30 per cent orthoclase, 30 per cent andesine, 7 per cent biotite, and 3 per cent muscovite. It is badly sericitized.

the principal reason why they are mapped as Tertiary by the U. S. Geological Survey (1933, map of United States) and by Philip King (1933, Pl. 1).

It may be thus concluded that while circumstantial evidence suggests early Tertiary intrusion, confirmative field evidence is not at hand.

AGE OF ROCKS IN NORTHERN TWO-THIRDS OF RUBY-EAST HUMBOLDT RANGE

The mountains north of Harrison Pass consist of sedimentary, metamorphic, and igneous rocks of complex structure and largely undetermined age. The previously named quartzite formation (Sharp, 1939b, p. 887) is undoubtedly the Lower Cambrian Prospect Mountain quartzite as suggested by Hill (1916, p. 57). Some of the overlying beds must be Cambrian, and the Ordovician Pogonip limestone has now been recognized just north of Harrison Pass. Devonian Devils Gate fossils have been found in the East Humboldt Range (Merriam, 1940, p. 43), and Carboniferous and Permian beds have been reported (King, 1878, p. 193, 205; Hill, 1916, p. 56).

High-grade metamorphic rocks, originally impure dolomite, limestone, shale, and sandstone, are exposed (Sharp, 1939b, p. 886-887) below the Lower Cambrian quartzite. These rocks are seemingly conformable with the quartzite, although the contact might be a low-angle thrust (Sharp, 1939b, Pl. 1, sec. EE′). They may be Lower Cambrian like the Black Rock limestone below the Brigham quartzite in Idaho (Anderson, 1928, p. 3-4; Ross, 1934, Fig. 2, p. 940) or like other Lower Cambrian strata of similar composition below Lower Cambrian quartzites (Nolan, 1929, p. 461-463; Noble, 1934, p. 175). They might correspond in part to the Proterozoic in Utah (Hinds, 1936, p. 73-92; Eardley and Hatch, 1940) or California (Knopf and Kirk, 1918, p. 9, 23-25; Maxson, 1935, p. 314; Noble, 1934, p. 174; Anderson, 1937, p. 61), for, as re-emphasized by Eardley and Hatch (1940, p. 837, 841), a noticeable break between probable Proterozoic rocks and the Lower Cambrian quartzite is lacking in many places.

King's (1877, p. 528, 537; 1878, p. 12, 62) dating of metamorphic rocks in the Ruby-East Humboldt Range as Archean has been accepted in part (Hague, 1892, p. 176; Spurr, 1903, p. 60). An Archean age was suggested because of the high-grade metamorphism and a supposed angular unconformity between Carboniferous limestone and the metamorphics. Many of these rocks are now known to be Cambrian or younger, and the supposed unconformity (King, 1878, p. 65) is a flat-lying thrust fault (Sharp, 1939b, p. 891).

In résumé, the sedimentary and metamorphic rocks in the northern Ruby-East Humboldt Range are largely and perhaps wholly Paleozoic. No indisputable evidence of pre-Cambrian rocks has been found, although strata beneath the Lower Cambrian Prospect Mountain quartzite may possibly be Proterozoic.

ROCK ALTERATION

SILICIFICATION

Silicified zones are common along the western flank of the southern Ruby Mountains and even more extensive to the north. Such bodies seem to have been mistaken by Hill (1916, p. 57, Pl. 3) for the Eureka quartzite.

Linear silicified zones, 10 to 25 feet wide, parallel the west base of the range south of Pearl Creek and also cross Toyn Creek. Elliptical silicified masses are exposed at the mouth of Walker Creek and just south of Cherry Springs. All these bodies consist of irregular aggregates of brecciated and healed crystalline and cryptocrystalline quartz containing country rock in various stages of replacement. The crystalline quartz is usually white, but the chalcedony may be white, bluish, or other colors depending upon the amount of undigested residue of country rock. Scattered drusy cavities and broadly curving striated shear surfaces are abundant. In places fractures in the silicified zones contain sulphides, and the whole mass is stained dark brown on weathered exposures.

Lindgren (1919, p. 155-158) attributes such bodies to replacement of limestone or dolomite by colloidal silica which later formed quartz or chalcedony. Essentially this view is adopted by Gilluly (1932, p. 97-101) and Nolan (1935, p. 93-94). Replacement of various rocks, including granite, along fracture zones has occurred in the southern Ruby Mountains. To the north silicified zones are well developed near igneous contacts, and although no igneous rocks are exposed near the silicified areas farther south such bodies may lie beneath the surface.

DOLOMITIZATION

Three types of dolomite are recognized in the southern Ruby Mountains: (1) widely distributed massive light-colored uniform dolomite, (2) dolomite of irregular structure and mixed colors and textures, and (3) brown massive coarse-grained dolomite.

The first type is represented by the Silurian Lone Mountain formation which is a light-colored massive dolomite of uniform texture, arrangement, and appearance. Silurian beds consist of this same type of dolomite over much of the Great Basin (Blackwelder, 1910, p. 527;

Richardson, 1913, p. 410; Lindgren and Loughlin, 1919, p. 103; Butler, Loughlin, and Heikes, 1920, p. 80; Mansfield, 1927, p. 58; Nolan, 1935, p. 17; Hazzard, 1937, p. 326; Merriam, 1940, p. 13).

The second type appears in the Devonian Nevada formation. It is distinguished by irregularities in color, texture, and arrangement. Most abundant is the mottled dolomite in which the contrasts are not nearly so marked as in the "mackerel" dolomite (Pl. 3, fig. 1). Most irregular are the beds composed of variously shaped masses of white dolomite complexly mixed with black dolomite (Pl. 3, fig. 2). The Nevada formation throughout this area is dolomitic, but elsewhere it is largely limestone (Merriam, 1940, p. 15).

The third dolomite type appears in several formations to the south. It is a medium- to coarse-grained massive light-brown or gray-brown dolomite that weathers dark brown and contains small wormlike intergrowths similar to the Bluebird dolomite (Lindgren and Loughlin, 1919, p. 28). The Ordovician Pogonip limestone, which elsewhere in the range is nondolomitic, and formations which ordinarily contain dolomite of a different type are here altered to the brown coarse-grained dolomite. The most distinctive feature of this type is its localization in a highly fractured area.

Most dolomitization is supposed to occur on the sea floor shortly after deposition of limy sediment (Van Tuyl, 1916; Twenhofel, 1932, p. 336-351). However, dolomites within the Great Basin have been attributed to a variety of processes, being both syngenetic and epigenetic at Pioche (Wheeler, 1940, p. 35-36); largely hydrothermal in the Goodsprings quadrangle (Hewett, 1931, p. 67); largely due to alteration in shallow water soon after deposition during subsequent stages of little or no sedimentation at Gold Hill (Nolan, 1935, p. 22); attributed to processes active during sedimentation and also during the pre-Mississippian land interval at Tintic (Lindgren and Loughlin, 1919, p. 103); caused by alteration immediately following deposition of lime mud in the Roberts Mountains region (Merriam, 1940, p. 22).

The southern Ruby Mountains dolomites are attributed to different modes of origin. The first type cannot be due to local conditions because of its uniformity and wide distribution. It was probably dolomitized on the sea floor shortly after deposition as suggested by Merriam (1940, p. 22). The second type was also probably formed on the sea floor presumably under local conditions of shallow water and intermittent deposition (Nolan, 1935, p. 22). The third type suggests a relationship between dolomitization and fractures. This dolomite is like that described by Hewett (1931, p. 57-67) and may likewise be attributed to changes caused by solutions circulating some time after the

sediments were lithified, perhaps contemporaneously with deformation. There is no proof that these solutions were hydrothermal as igneous rock is not exposed, although the complex structure and silicified zones may indicate an igneous body at depth.

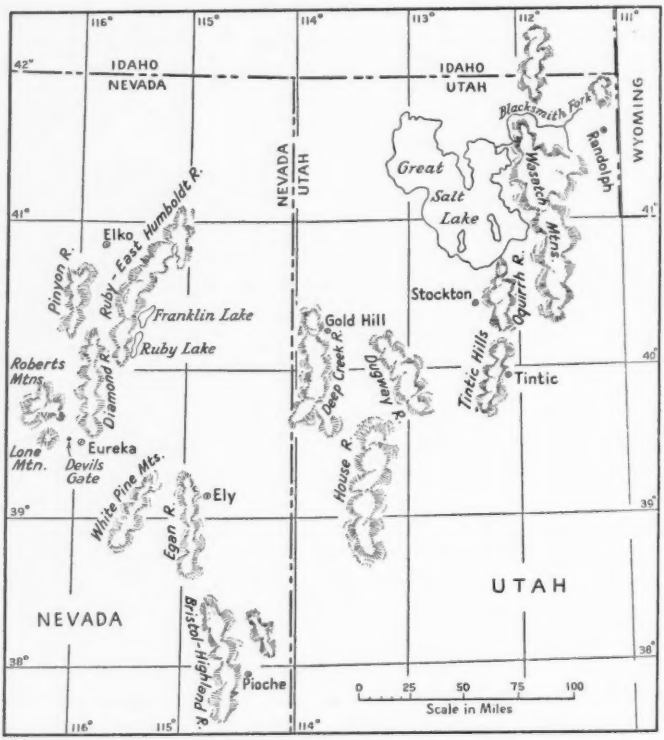

FIGURE 3.—*Map of eastern Great Basin localities*

PALEOGEOGRAPHIC INTERPRETATIONS

CAMBRIAN

Interpretations of Cambrian paleogeography are usually based on comparisons of thickness for the entire Cambrian system, but Gilluly (1932, p. 18) notes that such comparisons are best made only for the Middle Cambrian as the base of the Lower Cambrian is seldom exposed, and the top of the Upper Cambrian is often eroded. Middle Cambrian

thicknesses may be in error, however, because of inaccurate location of formation boundaries.

Table 1 shows that both the total Cambrian and the Middle Cambrian thicknesses in the southern Ruby Mountains occupy a medial position

TABLE 1.—*Thickness of eastern Great Basin Cambrian sections*

(Figures represent maximum thickness in feet)

Area	Cambrian	Middle Cambrian	Source of data
Gold Hill, Utah	12,050	4550	Nolan, 1935, p. 4-14
House Range, Utah	10,105	3420	Walcott, 1915, p. 175; Deiss, 1938, p. 1124-1147
Pioche, Nevada	9422	3500	Westgate and Knopf, 1932, p. 7; Wheeler, 1940, p. 12, 16
Tintic, Utah*	8846	1931	Lindgren and Loughlin, 1919, p. 22, 25-30
Ruby Mountains, Nevada	8050	3100	This report
Northern Wasatch Mountains, Utah	6885	4885	Richardson, 1913, p. 406-407; Deiss, 1938, p. 1108-1116
Eureka, Nevada†	6695	2100	Wheeler and Lemmon, 1939, p. 31, Fig. 3, p. 14
Oquirrh Range, Utah	2550	1200	Gilluly, 1932, p. 7, 18

* The sections at Tintic and Pioche may possibly contain some pre-Cambrian beds (Lindgren and Loughlin, 1919, p. 103; Wheeler, 1940, p. 16).

† The Eureka section is thinner than would ordinarily be expected, and this may be due to faulting.

in comparison with other eastern Great Basin sections. This demonstrates that these rocks were not deposited along the axis of the Cambrian Cordilleran geosyncline but somewhere on its flank. Gilluly (1932, p. 18) locates the Middle Cambrian geosynclinal axis about 100 miles east of the Ruby Mountains.

Thick Cambrian sections in southern Nevada (Nolan, 1929) and southeastern California[8] (Hazzard, 1937) contain a larger than average proportion of Lower Cambrian (Noble, 1934, p. 175) which suggests that initial invasion of the Cordilleran trough occurred through this area.

Identification of the Prospect Mountain quartzite in the Ruby Mountains extends the known area of Lower Cambrian farther north in this part of the Great Basin (Resser, 1934, p. 1) but furnishes no data on seaways around Montania in the northern Cordilleran geosyncline (Resser, 1934, p. 8-9; 1939, p. 6; Deiss, 1940, p. 788-789; 1941, p. 1095-1098).

[8] Hazzard (1937) first reported the Cambrian in the Nopah and Resting Springs mountains as 16,598 feet thick but later advised that the lower 4,050 feet is now believed to be Algonkian, thus reducing the Cambrian to 12,548 feet.

The Middle and Upper Cambrian beds of the southern Ruby Mountains become more argillaceous southward but even so are less argillaceous than beds of approximately the same age at Eureka. This may indicate shallower water or a source of sediment to the south or southwest during parts of the middle and late Cambrian. The less than average thickness of the Cambrian section at Eureka might be accounted for by such conditions.

ORDOVICIAN

The abundant intraformational conglomerate in Ordovician formations of the eastern Great Basin suggests changing conditions effective over a wide area. Furthermore, unconformities within the Ordovician (Nolan, 1935, p. 16; Richardson, 1913, p. 408; Kirk, 1933, p. 35, 39; Mansfield, 1927, p. 58, 173, 181) and at the top of the Pogonip in the southern Ruby Mountains attest to considerable unrest during that period.

Absence of the Eureka quartzite from a broad southward-projecting peninsulalike area in the northern Great Basin is reported by Kirk (1933, p. 32, 37-38), who infers that this may be due to nondeposition along the margin of a landmass. The southern Ruby Mountains are included in this area, and, if Kirk is right, the unconformity at the top of the Pogonip indicates uplift between Chazy and Black River time. However, there is no definite proof that the Eureka was not deposited in the Ruby Mountains and subsequently eroded. Under any circumstances upwarping of a broad area in the northeastern Great Basin between early middle Ordovician and the Silurian is indicated.

SILURIAN

Silurian formations throughout the eastern Great Basin are lithologically uniform but differ considerably in thickness. The thickest Silurian section yet reported is in the Roberts Mountains (Merriam, 1940, p. 19), 50 miles southwest of the Ruby Mountains, which contain a Silurian section less than half as thick. Merriam also reports a notable difference in thickness between Silurian sections in the Roberts Mountains and near-by Lone Mountain. Similar relations are described in other Great Basin areas. Merriam (1940, p. 19) attributes these differences largely to deposition, but they may also be due to post-Silurian erosion, for unconformities at the top of the Silurian are reported in several areas (Mansfield, 1927, p. 173; Nolan, 1928, p. 158; 1935, p. 18; Gilluly, 1932, p. 18-20; Ross, 1934, p. 998; Hazzard, 1937, p. 327-328). Differential uplift of parts of the Great Basin toward the close of the Silurian or later seems to be indicated. Nolan (1928, p. 158) suggests that uplifts of this nature may herald coming movements that produced notable changes by Carboniferous time.

DEVONIAN

Lack of paleontological evidence for the lower part of the Nevada formation in the southern Ruby Mountains may be due to the extensive dolomitization. However, emergence at the close of the Silurian followed by submergence in the Middle Devonian would be equally effective, and this sequence of events is indicated elsewhere (Mansfield, 1927, p. 173, 182; Gilluly, 1932, p. 20; Nolan, 1935, p. 18).

The distinct lithologic difference between the Middle and Upper Devonian parts of the Devils Gate in this area contrasts with the more uniform character of the formation at the type locality (Merriam, 1940, p. 16). Considerable clastic material in the upper Devils Gate and chert-pebble conglomerate at its base probably indicate initial phases of movements that notably affected the paleogeography of the Great Basin by Carboniferous time (Nolan, 1928). Apparently these changes were effective earlier in the Ruby Mountains area than in the Roberts Mountains region.

CARBONIFEROUS

Shale and quartzite at the top of the Mississippian(?) and the siliceous conglomerate and quartzite in the undifferentiated Carboniferous are evidence of a considerable change from the earlier Paleozoic. Conditions during the Carboniferous were variable over much of the Great Basin as shown by interbedded conglomerate and limestone.

The source of the pebbles in the conglomerate is unknown. Their universal siliceous composition suggests elimination of less resistant rock types by previous deep weathering or by extended corrasion and transportation. The pebbles may be of a second or even third generation, and the original source may have been multiple.

STRUCTURE

INTERNAL STRUCTURE OF THE MOUNTAINS

Rattlesnake Mountain dome.—This structure, exposed on Brown Creek (Pl. 1), is the only fold of any size south of Harrison Pass. It is an asymmetrical dome trending roughly north with the east limb crenulated and locally overturned and the western limb partly cut off by the fault bounding the west base of the range. The fold has a core of granite which intrudes the deformed beds. Eastward overturning suggests pressure from the west. The deformation is not dated more precisely than late or post-Paleozoic.

Thrust faults.—A broad thrust sheet of Carboniferous rocks resting on Cambrian and Ordovician strata composes the west flank of the range between Cass House and Sherman creeks. Outliers are scattered along

the base of the range south to Overland Pass where a large section of this thrust sheet is downfaulted. The dip of the thrust plane as measured in field exposures is 16°-20° W. Locally it must be steeper, for outcrop relations cannot be explained satisfactorily unless the fault plane is uneven. Part of this unevenness is probably due to differential uplift of the mountain block, and the original westward dip of the plane was probably 5° to 6° less owing to westward tilting during that uplift.

The dip of the fault plane and local overturning indicate displacement from the west. Assuming simple structure and average dips, thrusting of Carboniferous onto Cambrian requires a minimum displacement of 40,000 feet. The nearest known exposure of Carboniferous beds of the type composing the thrust sheet is in the northern part of the Diamond Range and Diamond Hills, 10 miles west (Fig. 3). An eastward displacement of 7 to 10 miles along the thrust is indicated.

Just to the east and roughly parallel with the low-angle thrust is a steep thrust that dips 60°-70° W. This fault extends from the North Fork of Lindsay Creek, where it passes into a sharp flexure, to a point south of Sherman Creek where it passes under the low-angle thrust. A small outlier of the low-angle thrust sheet extends across the high-angle fault north of Mitchell Creek and is further proof that the high-angle fault is the older, although both are probably related to the same period of deformation. A maximum dip slip of 3500 to 4000 feet along the steep fault has brought Cambrian into contact with Ordovician and Silurian beds. Other small high- and low-angle thrust faults in this area reflect the characteristics of these larger faults.

Local field data permit the thrusting to be dated only as post-Carboniferous, but analogies with similar thrusts in neighboring regions (Butler, Loughlin, and Heikes, 1920, p. 102; Ferguson, 1929, p. 117; Hewett, 1931, p. 55; Billingsley, 1933, p. 22) suggest that it may be early Tertiary.

Normal faults.—Only the normal faults within the mountain block are described here. They are numerous only in the south where a predominant group with a northeast trend and the downthrown side on the southeast is crossed by an east-west group with downthrow on the south. Dips measured on actual exposures of the fault planes and zones are 60° to 70°, although some appear to be steeper. Displacements are between 1000 and 4000 feet.

Two subparallel normal faults near the east base of the range south of Flynn Springs dip 70° E. A downthrow of 1000 to 1200 feet of the eastern blocks has brought Mississippian(?) into contact with Middle Devonian. A normal fault at the east base just south of Cave Creek

dips 80° E., and upper Devonian beds have been dropped down on the east into contact with Middle Devonian.

These normal faults are all attributed to the same period of deformation because of similarity in trend, dip, and displacement. They displace the low-angle thrust in Overland Pass and are believed to be older than the faults bounding the mountain block because they lack fault scarps along their traces.

BOUNDARY STRUCTURE

General statement.—Evidence showing that the northern two-thirds of the Ruby-East Humboldt Range is a westward-tilted horst has already been presented (Sharp, 1939b). The northern part of the southern Ruby Mountains continues this structure, but farther south the west boundary fault dies out, and the range becomes simply a westward-tilted fault block.

East boundary fault.—This fault is buried beneath the alluvium of Ruby Valley. However, the abruptness (Fig. 1) and linear trend of the east face, the discordance between the range front and the internal structure (Pl. 1), and the line of springs along the base are evidence of a fault at the east base of these mountains. The trend of this fault is north to north-northeast, and from information obtained farther north (Sharp, 1939b, p. 898) it is assumed to be a normal fault with the downthrow to the east. The topographic relief caused by fault displacement is 4500 feet, and south of Harrison Pass a forelying Carboniferous block has been dropped into contact with Silurian strata composing the mountain block by a dip-slip displacement of not less than 6500 feet. Steepening of bedding along the east base of the range near Cave Creek is due largely to drag on a smaller normal fault already described. The present range face has been eroded back ¼ to ½ mile from the postulated position of the east boundary fault.

West boundary fault.—This fault approximately parallels the east boundary fault and extends south nearly to Sherman Creek (Pl. 1) where the mountain block passes from a tilted horst to a simple westward-tilted block. Truncation of the internal structure, particularly at Brown Creek, and a small in-faulted block of eastward-dipping Miocene beds between Mitchell and Sherman creeks are attributed to the west boundary fault. No exposures of the fault plane have been observed, but analogies drawn from the northern part of the range indicate that it is a normal fault with downthrow to the west. The displacement appears to be about 1000 to 2000 feet and decreases southward.

Age of boundary faults.—Displacements along the boundary faults of the Ruby-East Humboldt Range occurred at intervals from the Miocene to late Pleistocene or Recent (Sharp, 1939b, p. 902-905), but there is little direct evidence for this in the southern Ruby Mountains. Post-Humboldt movement is established by displacement of Miocene Humboldt beds near Sherman Creek. The Quaternary fanglomerate is attributed to movement on the east boundary fault. Dissection of this deposit is also probably due to a subsequent displacement along the east boundary fault and indicates an absolute upward movement of the mountain block. The fanglomerate is not displaced by the west boundary fault so no subsequent movement occurred along the southernmost part of this fault. The asymmetry of the range and the abruptness of the east face suggest a greater and in part more recent displacement on the east side.

GEOMORPHOLOGY

GEOMORPHIC FEATURES OF THE MOUNTAIN BLOCK

Glaciation.—Two substages of Wisconsin glaciation, the Lamoille (early Wisconsin = Iowan) and the Angel Lake (later Wisconsin), have been recognized (Blackwelder, 1931, p. 910-911; 1934, p. 218-219; Sharp, 1938). Evidence of glaciation in this area consists of cirques, U-shaped canyons, moraines, and outwash deposits. Fresh well-developed cirques are located on the flanks of high peaks (Fig. 1; Pl. 5, fig. 2) as far south as Sherman Mountain. Morainal deposits consist of lateral ridges, terminal and recessional loops, and irregular ground moraine. Exceptionally large moraines, 600 to 700 feet high, choke the canyons east of Pearl Peak (Pl. 5, fig. 3). Outwash materials have been deposited in some canyons and compose exceptionally large steep fans at the east base of the range.

Glaciers of the Lamoille substage were the larger, and the longest glacier in this part of the range extended 2½ miles down the South Fork of Lindsay Creek. The lowest altitude, 6900 feet, was attained by a glacier in a large canyon on the northeast flank of Pearl Peak. The southernmost glacier formed on the north flank of Sherman Mountain where altitude, catchment area, and exposure were particularly favorable.

Wisconsin orographic snow line[9] was close to 10,000 feet, which is several hundred feet higher than the average for the northernmost part of the range. The Wisconsin snow line rose southward approximately 200 to 300 feet in every 100 miles, and this checks well with the gradient of 300 feet per 100 miles calculated for the distance between the Ruby Mountains and San Francisco Mountain in northern Arizona. However,

[9] *See* L. L. Ray (1940, p. 1910-1911) for a discussion of relative significance of orographic and climatic snow lines.

Klute (1928, p. 75) gave 500 feet per 100 miles for the climatic snow line gradient in the Rocky Mountains, and Kessler (1925, p. 171-172) cited 580 feet per 100 miles for Germany. Local factors are obviously of considerable importance.

Miscellaneous features.—The Ruby-East Humboldt Range is a youthful Basin Range whose outline and form have been determined by block faulting. Extensive dissection has almost completely destroyed geomorphic forms which preceded development of the range as an orographic unit or formed during pauses in uplift. An open valley at the head of Mitchell Creek and broad range-crest flats on Sherman and Willow creeks are possibly remnants of the middle or late Pliocene open-valley stage recognized farther north (Sharp, 1940, p. 342-343). At the heads of Willow, Sherman, and Mitchell creeks the drainage divide is east of the range crest, thus duplicating a relation previously reported to the north and attributed to a complex interplay of Humboldt River rejuvenations and uplifts of the range (Sharp, 1940, p. 343-346).

Harrison Pass, with a relatively low summit at 7248 feet, has been developed upon a coarse-grained granitic stock. This granite succumbs readily to granular disintegration and subsequent erosion by running water. Overland Pass has been formed by relatively rapid erosion of a badly fractured and faulted area.

Within the mountains subterranean caverns and passages have been formed in massive limestone by solution. Cave Creek emerges from one of these passages, and its large uniform volume indicates that the water comes from a large drainage area. Undoubtedly some of the Cave Creek water has been derived from west of the range crest having been led eastward into the Cave Creek drainage system by the eastward-dipping sedimentary beds.

Small stone nets (*Polygonboden*) have been produced by frost action on the range-crest flat east of Sherman Mountain. The small size and freshness of these structures suggest that they are of recent origin (Högbom, 1914, p. 309-315; Elton, 1927; Antevs, 1932, p. 49-53; Denny, 1940).

GEOMORPHIC FEATURES OF THE RANGE FLANKS

East flank.—A narrow alluvium-mantled pediment is believed to extend along the east base of the range, for such a pediment is exposed farther north. Lack of dissection in Ruby Valley, a closed basin, makes direct confirmation impossible. This pediment was formed with respect to a rising base level, hence its burial.

Shore lines of a late Pleistocene lake are located around the edges of Ruby Valley. The highest shore line is close to 6200 feet. At this stage

the lake was nearly 300 feet deep, and its waves lapped directly against the face of the range between Cave Creek and Flynn Springs. Elsewhere, lake cliffs are cut in gravels. Beaches of reworked water-worn pebbles are located at the base of the lake cliffs or form linear embankments. East of Flynn Springs a large spit extends northeastward part way across the mouth of an embayment indicating a shore current in that direction. This lake had no direct outlet but at its highest stage joined with Franklin Lake to the north, and this composite body may have had an outlet to the east.

West flank.—Remnants of erosion surfaces 1, 2, 3, and 4 recognized along the west flank farther north (Sharp, 1940, p. 346-353) are preserved near Toyn and Corral creeks. Southward the range is bordered by a broad gravel-mantled pediment that truncates Miocene Humboldt beds and in minor part the harder rocks of the mountain block. This surface may be the time equivalent of several surfaces recognized farther north, for the rejuvenations of the Humboldt River drainage previously described (Sharp, 1940, p. 357-359) are not represented here. Possibly at this time the streams in the southernmost Ruby Mountains drained southward into Newark Valley east of the Diamond Range (Fig. 3) and not northward via Huntington Creek to the Humboldt River as at present.

To the south remnants of fanglomerate rest upon an erosion surface that truncates bedrock composing the mountain block. Dissection of this surface is attributed to uplift along the east boundary fault accompanied by westward tilting. The depth of dissection, 400 feet, suggests that it is related to surface 1 recognized farther north or to the open-valley stage in the mountain block.

SUMMARY

The southern third of the Ruby Mountains contains a Paleozoic section more than 20,000 feet thick composed of the following units:

	Feet
Lower Cambrian—Prospect Mountain quartzite	1400
Middle Cambrian	3100
Upper Cambrian	3550
Ordovician—Pogonip limestone	3650
Silurian—Lone Mountain formation	1350
Middle Devonian—Nevada formation	1900
Middle and Upper Devonian Devils Gate formation	1200
Mississippian(?)	300
Undifferentiated Carboniferous	4000 to 10,000

The strata from Cambrian to Mississippian(?) comprise a continuous sequence 16,450 feet thick from which the undifferentiated Carboniferous beds are separated by faults. Exceptionally simple structure and good exposure make this section worthy of more detailed study. The Miocene

Humboldt formation, a Quaternary fanglomerate, glacial drift, and gravels are also described.

Stocks and dikes of granitic rock intrude the Paleozoic strata in the north and can be dated only as post-Paleozoic and pre-Miocene, although by analogy with neighboring areas they may be early Tertiary.

The Lower Cambrian Prospect Mountain quartzite is widely distributed in the northern two-thirds of the Ruby-East Humboldt Range, and other Cambrian, Ordovician, Silurian, Devonian, and later Paleozoic rocks are also represented. Strata below the Prospect Mountain may be Lower Cambrian or possibly Algonkian, but not Archean.

Cambrian and Ordovician beds near igneous bodies are metamorphosed to hornfels, schist, granulite, and related metamorphic types. The various dolomites described are attributed chiefly to alteration on the sea floor shortly after deposition and locally, in highly fractured areas, to later circulating solutions of unknown origin.

Previous interpretations of Paleozoic paleogeography in the eastern Great Basin are largely confirmed in this area. An unconformity at the top of the Pogonip and the absence of the Eureka quartzite are attributed to upwarp of a broad area in the northeastern Great Basin between early middle Ordovician and the Silurian. Wide variation in thickness of Silurian sections is attributed in part to post-Silurian erosion. The forces which produced great changes in paleogeographic relations by Carboniferous time were active in this area by late Devonian.

The internal structure is a simple eastward-dipping homocline complicated by igneous stocks and dikes in the north, the Rattlesnake Mountain dome at Brown Creek, low- and high-angle thrusts along the west flank, and a complex of high-angle normal faults in the south. The low-angle thrust on the west flank dips 16 to 20° W., and an eastward displacement of 7 to 10 miles has put undifferentiated Carboniferous strata on top of Cambrian and Ordovician beds. By analogy with neighboring areas the thrusting may be early Tertiary. A northeast trend and consistent downthrow to the southeast dominate the complex of high-angle normal faults. These faults are probably Tertiary but are older than the normal faults bounding the range.

The northern part of the southern Ruby Mountains is a westward-tilted horst. The maximum displacement measured on the east boundary fault is 6500 feet, and it is estimated at 1000 to 2000 feet on the west boundary fault. Displacement on this latter fault decreases southward, and the fault dies out so that the southernmost part of the range is a simple westward-tilted fault block. Displacement on these faults occurred repeatedly from Miocene to sometime in the Pleistocene, and uplift has been greater and in part more recent along the east boundary fault.

Features of an early Wisconsin (Iowan) and a later Wisconsin glaciation indicate a Wisconsin orographic snow line at about 10,000 feet in this part of the range. Calculations show that this snow line rose southward 200 to 300 feet in every 100 miles in this part of the Great Basin.

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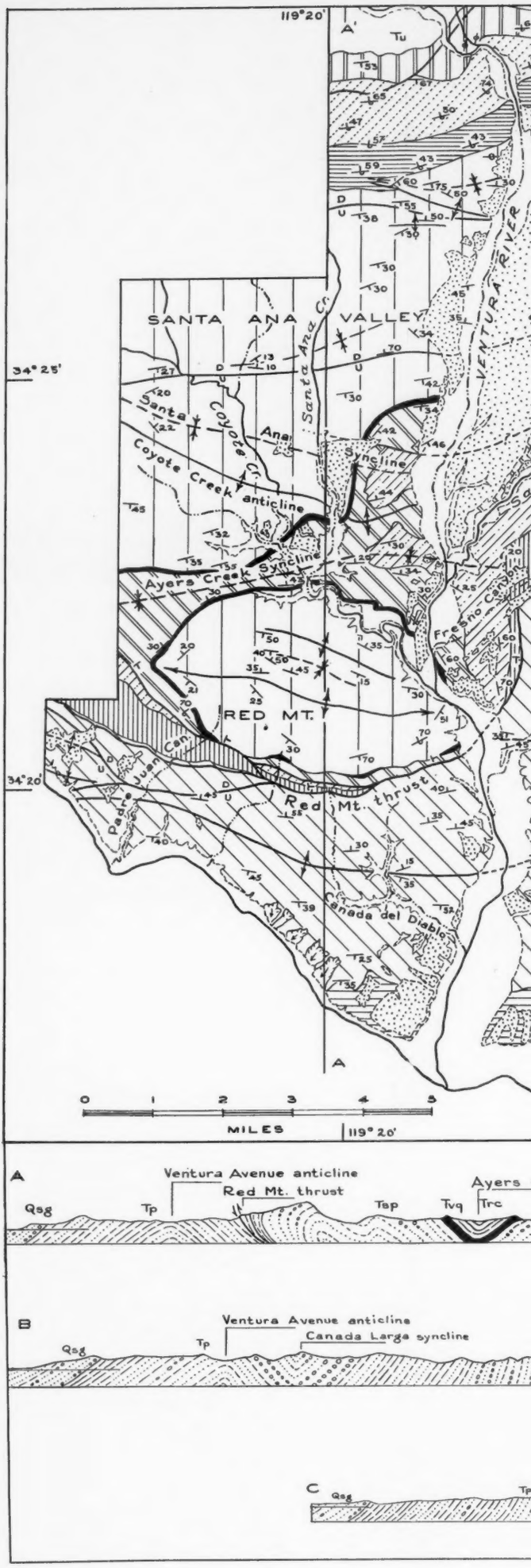
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SANTA ANA VALLEY
VENTURA RIVER
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34°25'
Santa Ana Syncline
Coyote Cr.
Coyote Creek anticline
Ayers Creek Syncline
Fresno Canyon
RED MT.
Pole Juan Can.
34°20'
Red Mt. thrust
Canada del Diablo
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MILES
119°20'
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Ventura Avenue anticline
Red Mt. thrust
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Canada Larga syncline
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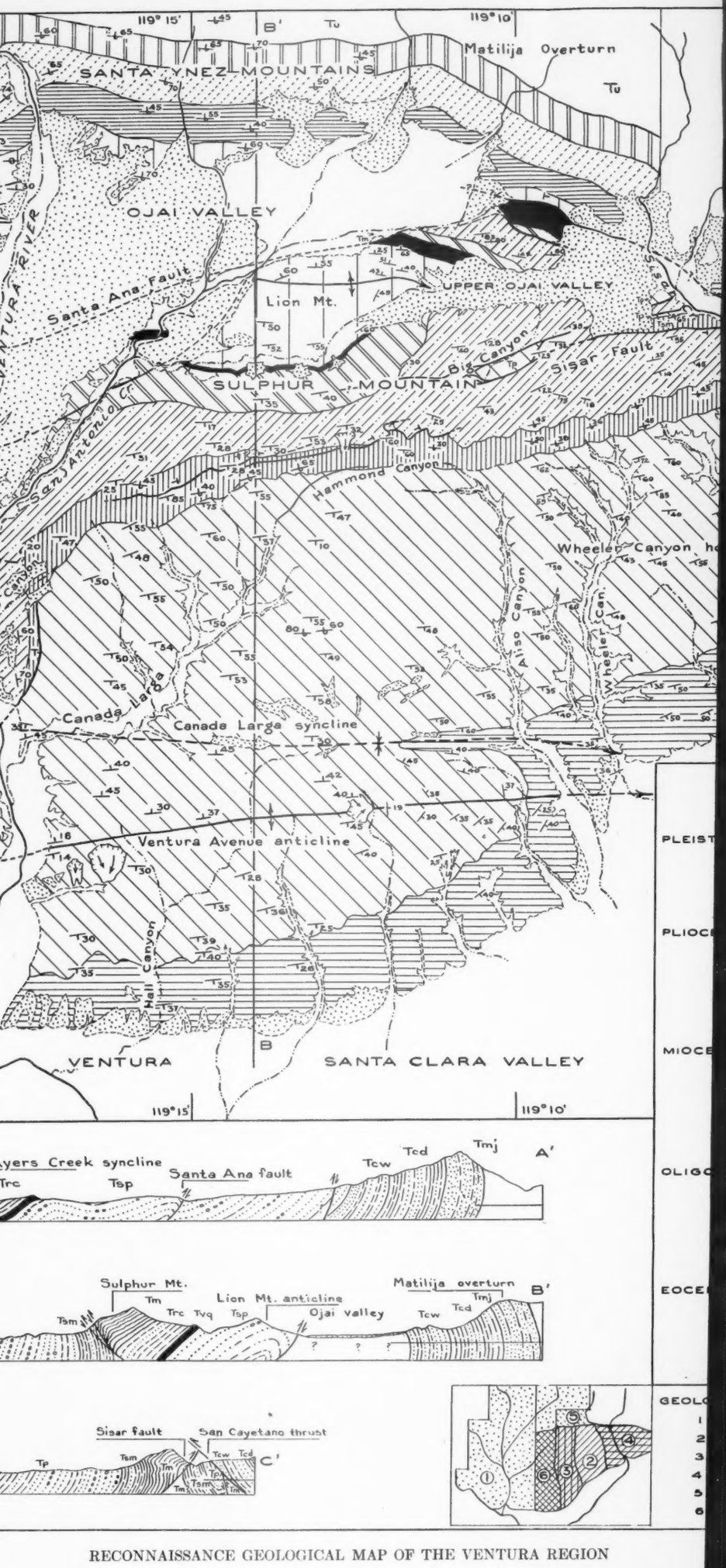
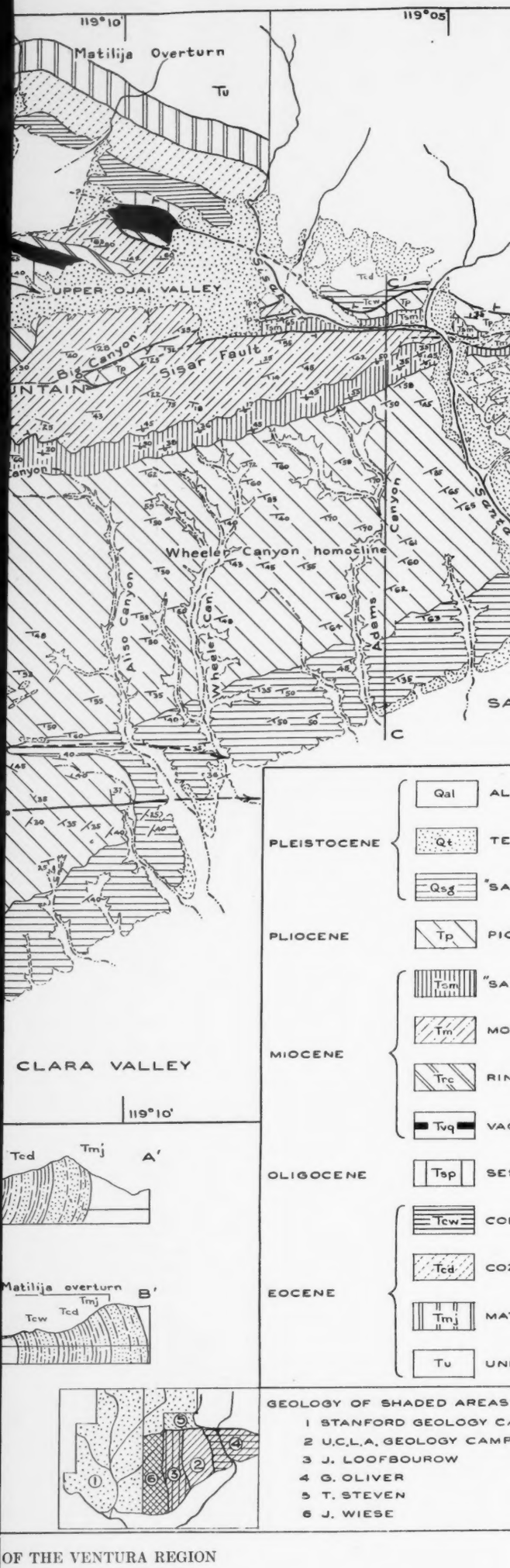


RECONNAISSANCE GEOLOGICAL MAP OF THE VENTURA REGION



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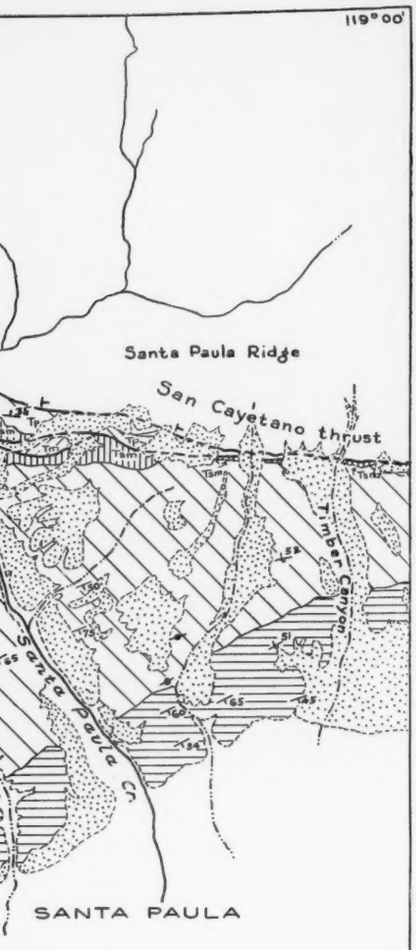

ALLUVIUM

TERRACES & OLDER ALLUVIUM

"SAUGUS"

PICO

"SANTA MARGARITA"

MODELO

RINCON

VAQUEROS

SESPE

COLDWATER

COZY DELL

MATILIJA

UNDIFFERENTIATED

AREAS BY:
OGY CAMP, 1937-1938
CAMP, 1941

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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GEOMORPHOLOGY OF THE VENTURA REGION, CALIFORNIA

BY WILLIAM C. PUTNAM

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ABSTRACT

In the Ventura region approximately 47,000 feet of exposed Cenozoic sedimentary rocks provide a nearly complete record of the later events in the sedimentational history of the southern Coast Ranges. The Pleistocene history may be divided into four episodes: (1) the deposition of 6000 feet of marine and terrestrial strata; (2) folding and faulting of all the formations; (3) stream erosion which carved a surface of late maturity, and; (4) intermittent regional uplift which started the present erosion cycle, now in the stage of early maturity. Differential erosion has conspicuously controlled the topography. Active faulting and warping occur widely. The drainage diversion of the Upper Ojai Valley by Santa Paula Creek is a large-scale example of stream capture, and numerous other examples of drainage adjustment have taken place. Late Pleistocene uplift is recorded by terraces in the Ventura River valley, and by marine terraces bordering the Pacific Ocean. Both series of terraces are warped, and both reach their maximum elevation where the rocks are intensely deformed.

INTRODUCTION

INTRODUCTORY STATEMENT

The Ventura region provides a nearly complete record of the Cenozoic history of coastal southern California. Cenozoic strata in the Ventura basin are 47,000 feet thick; of these approximately 4000 to 6000 feet accumulated as marine and terrestrial deposits in the early Pleistocene. All were deformed in the medial Pleistocene, and beveled by an erosion surface of late maturity. Differential vertical uplift, responsible for the river and marine terraces, rejuvenated the drainage system and initiated the present erosion cycle, now in the stage of maturity.

Ventura is 69 miles northwest of Los Angeles (Fig. 1), and the region here discussed is included in the boundaries of the Ventura and Santa Paula quadrangles of the U. S. Geological Survey. The writer first be-

came interested in 1930 in geomorphic problems near Ventura and continued field work there during parts of the summers of 1932, 1933, and 1935 to 1941. The emphasis of this paper is on the origin of the Ventura landscape, the evolution of the drainage pattern, and the deformation of the terraces.

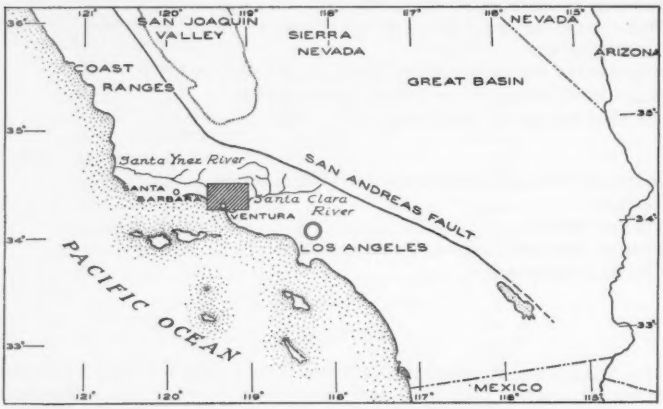

FIGURE 1.—*Index map*

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PHYSIOGRAPHIC PROVINCES

Approximately one fourth of the Ventura quadrangle is covered by the Pacific Ocean. Most of the land surface of the two quadrangles is hilly or mountainous. The only important lowlands are the Santa Clara Valley, the Ventura River valley, and the Ojai intermont basin.

These topographic differences are the basis for subdividing the region into four physiographic provinces (Fig. 2):

(1) Santa Ynez Mountains: the east-trending mountain chain on the north border of the area. The maximum altitude is 6351 feet.
(2) Ojai Lowland: the intermont basin between the Santa Ynez Mountains and Sulphur and Red mountains and including the Santa Ana, Ojai, and Upper Ojai valleys. The average altitude of the Ojai Valley is 750 feet.
(3) The Sulphur Mountain Upland: the discontinuous highland through the center of the area formed by Sulphur, Red, and Rincon mountains. The average altitude of Sulphur Mountain is 2000 feet.
(4) Coastal Hills: the low, hilly region adjacent to the coast and extending inland parallel to the Santa Clara River to east of Santa Paula. The maximum elevation is 1950 feet.
(5) Santa Clara Valley: the flood plain of the Santa Clara River at the south boundary of the region.

CLIMATE

The climate is subtropical, hot-dry (Zierer, 1932, p. 28-29). The winter maximum of rainfall is more significant than the total amount. The range for the past 10 years was 6 to 25 inches, and the average about 15. Erosion characteristic of arid regions might be expected. The smooth slopes and rounded summits reflect the seasonal rainfall, as pointed out by Bryan (1923, p. 20) and others (Woodring, 1932, p. 8; Stearns, 1930, p. 20) for similar California areas. According to Bryan, in the Sacramento Valley:

"In August and September the soil is dry, hard, and commonly cracked to considerable depths. With the first rains, which usually come in October, grass springs up, either from seed or perennial rootstocks. With the usual mild winter temperatures growth of grass and other vegetation continues through the winter and culminates in the spring.

"The effect of this vegetative cover on erosive processes is striking, for the surface of the ground is protected during the five winter months of rainfall, and the erosion takes place under conditions similar to those of a humid climate. During the summer when the grass dries up and the surface of the ground is least protected, so little rain falls that no erosion takes place. In effect, the seasonal rainfall of the Sacramento Valley gives the topographic results of twice the amount. . . .

"The soils, however, because of the freedom from plant growth and of thorough oxidation in the dry summer season, are similar to other arid soils in containing undecomposed feldspars and other minerals, and in high nitrate content. It seems probable that erosion being operative for half the year only, is only half as rapid as in a humid climate."

STRATIGRAPHY

GENERAL CONSIDERATIONS

The rocks are conglomerates, sandstones, shales, and thin-bedded limestones, and, except for the Sespe and upper "Saugus" strata and the terrace gravels and alluvium, are marine. The formations cropping out in the Ventura region are named on the columnar section (Fig. 3), and their areal extent and distribution is shown on the geologic map (Pl. 1).

Each formation has a distinctive topographic expression, and this aspect of stratigraphy is the proper concern of the paper. The most

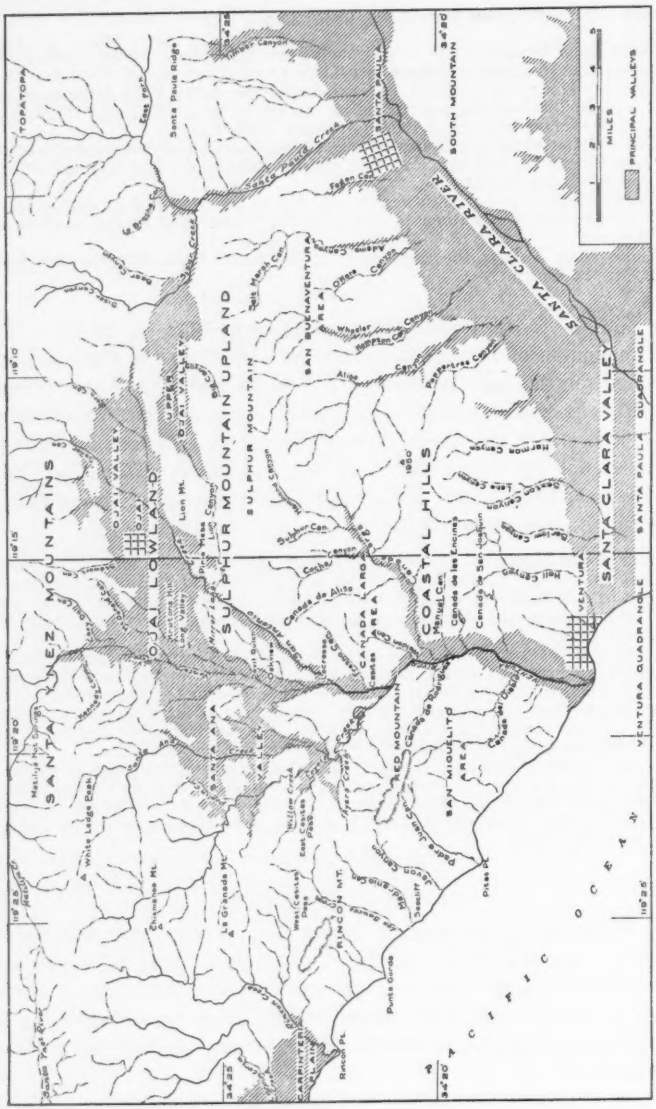

FIGURE 2.—*Geographic map*

resistant rocks are the Matilija sandstones. These crop out on the southern slope of the Santa Ynez Mountains as steep cliffs where flat-lying, or as impressive hogbacks where steeply dipping. The white Coldwater sandstone makes three subordinate ridges at the base of the

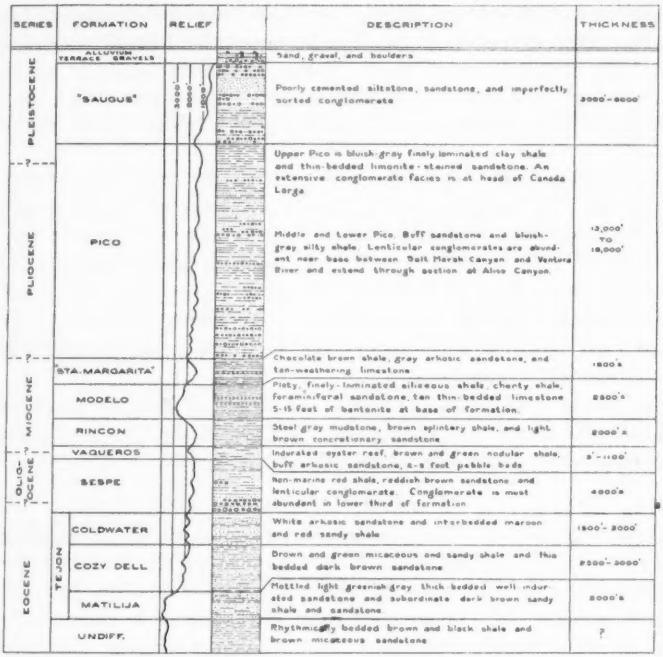
SERIES
FORMATION
RELIEF
DESCRIPTION
THICKNESS
ALLUVIUM TERRACE GRAVELS
Sand, gravel, and boulders
PLEISTOCENE
"SAUGUS"
Poorly cemented siltstone, sandstone, and imperfectly sorted conglomerate
2000'-6000'
PLIOCENE
PICO
Upper Pico is bluish-gray finely laminated clay shale and thin-bedded limonite-stained sandstone. An extensive conglomerate facies is at head of Canada Larga
Middle and lower Pico. Buff sandstone and bluish-gray silty shale. Lenticular conglomerates are abundant near base between Salt Marsh Canyon and Ventura River and extend through section at Alisa Canyon.
13,000' TO 18,000'
MIOCENE
STA. MARGARITA
Chocolate brown shale, gray arkosic sandstone, and tan-weathering limestone
1800'±
MODELO
Platy, finely-laminated siliceous shale, chonty shale, foraminiferal sandstone, tan thin-bedded limestone 5-15 feet of bentonite at base of formation.
2500'±
RINCON
Steel gray mudstone, brown splintery shale, and light brown concretionary sandstone
2000'±
OLIGOCENE
VAQUEROS
Indurated oyster reef, brown and green nodular shale, buff arkosic sandstone, 6-8 feet pebble beds
5'-1100'
SESPE
Non-marine red shale, reddish brown sandstone and lenticular conglomerate. Conglomerate is most abundant in lower third of formation
4000'±
EOCENE
TEJON
COLDWATER
White arkosic sandstone and interbedded maroon and red sandy shale
1500'-2000'
COZY DELL
Brown and green micaceous and sandy shale and thin bedded dark brown sandstone
2500'-3000'
MATILIJA
Mottled light greenish-gray thick bedded well indurated sandstone and subordinate dark brown sandy shale and sandstone
2000'±
UNDIFF.
Rhythmically bedded brown and black shale and brown micaceous sandstone
?

FIGURE 3.—*Generalized columnar section*

Santa Ynez Mountains, and is usually separated from the Matilija hogback by a lowland in the Cozy Dell shale.

Broad valleys and low hills are eroded in the red Sespe shales and sandstones where these rocks are gently folded, as in the Santa Ana and Ojai valleys. Hogbacks with bare sandstone dip slopes are prominent west of the Santa Ana Valley where the Sespe is part of the south-dipping homocline. The anticlinal structure of Red and Lion mountains (Pl. 1) is made apparent by Sespe sandstone beds that crop out in resistant ribs.

In the dry climate of southern California the thin oyster reef of the Vaqueros is a resistant rock. The reef is highest where the lime content

is greatest, and makes two opposite-facing cuestas on the limbs of the Ayers Creek syncline (Pl. 1, Section A-A′). The overlying Rincon shales are weak and make a lowland, best developed north of Sulphur Mountain at Lion Canyon and along the axis of the Ayers Creek syncline west of the Ventura River.

Sulphur and Rincon mountains owe their superior elevation to the fact that they are underlain by the siliceous shale of the Modelo formation. The closely jointed, brown "Santa Margarita" shale is markedly less resistant, and subsequent streams have etched out strike canyons with the Sulphur Mountain escarpment in the Modelo forming the north side.

The conglomerate, sandstone, and shale of the Pico are resistant in that order. The conglomerate is largely restricted to the lower part of the formation, except west of Aliso Canyon where it extends through the section. The steeply dipping conglomerate makes hogbacks, best developed at the head of Wheeler Canyon, and also holds up the highest peak (1950 feet) in the Coastal Hills. Clay shales, particularly in the upper Pico, erode to wide valleys and low hills whose rounded slopes are scarred with earthflows.

The poorly consolidated "Saugus" sands and gravels are more resistant than the upper Pico clay shale. Valleys in the "Saugus" are narrower and their walls steeper than in the Pico.

TERRACE GRAVEL

General statement.—This section is concerned with the stratigraphy and paleontologic content of the terrace gravels, rather than their physiographic history. The terraces are of two kinds: marine terraces and stream terraces. The latter are strewn with rounded boulders and cobbles, chiefly of Eocene sandstones, but including some from other formations. Gravels on higher terraces are more oxidized than on lower benches. The matrix weathers deep reddish brown and the surface of boulders is stained. The average diameter of large boulders on the lowest terrace decreases from 2 to 4 feet in the Ojai Valley to less than 1 foot near the river mouth, although boulders 5 feet by 6 feet are found on high terraces near the coast.

Few boulders on marine terraces exceed 1 foot, but some attain 4 or 5 feet in diameter. The matrix contains more silty sand than the river terraces and is generally oxidized light yellow brown. Many of the cobbles and boulders are stained deep reddish brown. The thin marine and littoral deposit on the terrace floor is buried under a thick cover of terrestrial material washed from the abandoned sea cliffs or higher hills.

Vertebrate fossils.—The stream gravels contain vertebrate remains. Teeth of *Equus occidentalis* have been recovered at a number of localities, and *Elephas imperator* teeth were found in tilted terrace gravels at the mouth of Barlow Canyon. According to Kellogg (1922, p. 61), Bowers (1889, p. 391) reported the finding of *Eumetopias stellari* (=*jubata*) associated with *Mastodon shephardi, Holomeniscus californicus, Equus occidentalis,* and *Eschrichtius davidsoni* in street cuts in Ventura.

Marine invertebrate fossils.—Fossil mollusk shells are found in gravelly sand overlying the wave-abraded surface of marine terraces. Collections were made at two localities, and the fauna recovered there is listed in Table 1. Except for *Cardita monilicosta,* all species are living, and all include the Santa Barbara Channel in their modern range. Inspection of the ranges shown on the faunal list indicates that the majority are cool water forms. A similar assemblage would be expected in the vicinity of Cape San Martin, 3° farther north (Schenck, personal communication, 1941).

TABLE 1.—*Fossils from Ventura marine terraces*

	Locality		Range in
PELECYPODA	1	2	Latitude*
Apolymetis biangulata (Carpenter)	x	x	33-34:33
Cardita monilicosta Gabb	x		23-34:28
Chione fluctifraga (Sowerby)	x	x	23-34:28
Clinocardium nuttallii (Conrad)	x		33-60:46
Cryptomya californica (Conrad)		x	11-59:35
Glycymeris corteziana Dall	x	x	33-55:44
Macoma inconspicua (Broderip and Sowerby)	x		33-72:52
M. nasuta (Conrad)	x	x	28-60:44
M. secta (Conrad)	x		23-49:36
Nuculana taphria (Dall)		x	30-38:34
Pachydesma crassatelloides (Conrad)		x	19-37:28
Pholadidea penita (Conrad)		x	33-56:44
Platyodon cancellatus (Conrad)	x	x	33-38:35
Protothaca staminea (Conrad)	x	x	23-73:48
Saxidomis nuttalli (Conrad)	x	x	30-41:35
Schizothaerus nuttalli (Conrad)	x	x	28-38:33
Spisula hemphilli (Dall)		x	32-34:33
Zirfaea pilsbryi Lowe	x		28-56:42
SCAPHOPODA			
Cadulus fusiformis Pilsbry and Sharp	x		23-37:30
GASTROPODA			
Acanthina spirata (Blainville) ss	x		30-34:32
Antiplanes perversa (Gabb)	x		30-55:42
Borsonella bartschi (Arnold)	x	x	33-34:33
Calliostoma canaliculatum (Martyn)	x		33-57:45
Calyptraea contorta (Carpenter)		x	23-34:28
Clathrodrillia incisa (Carpenter)		x	33-48:41
C. incisa ophioderma Dall	x		27-37:32

TABLE 1.—*Fossils from Ventura marine terraces*—Continued

	Locality 1	Locality 2	Range in Latitude*
Conus californicus Hinds	x		26-38:32
Crepidula adunca Sowerby	x		23-48:35
C. nummaria Gould	x		23-65:44
Diodora aspera (Eschscholtz)	x		24-60:42
Epitonium tinctum (Carpenter)	x		23-49:36
Euspira lewisii (Gould)		x	33-51:42
Fusinus kobelti monksae (Dall)	x	x	26-50:38
Glyphostoma conradiana (Gabb)		x	34:34
Hipponix antiquatus (Linnaeus)	x		−1-42:20
Homalopoma carpenteri (Pilsbry)	x		30-57:43
Mangelia hecetae Dall	x		33-48:40
Mitrella carinata (Hinds)	x	x	23-38:30
Nassarius cooperi (Forbes)	x	x	33-48:40
N. fossatus (Gould)	x	x	27-49:38
N. mendicus (Gould)	x	x	25-55:40
N. perpinguis (Hinds)	x	x	28-48:38
Olivella biplicata (Sowerby)	x	x	25-49:37
O. pedroana (Conrad)	x	x	23-55:39
Neverita reclusiana (Deshayes)	x		21-42:31
Pseudomelatoma moesta (Carpenter)	x		27-37:32
P. penicillata (Carpenter)	x		28:28
Tegula montereyi (Fischer)	x		34-38:36
Thais lamellosa (Gmelin)	x		34-65:49
Tritonalia lurida (Middendorff)	x		34-55:44
T. poulsoni (Nuttall *in* Carpenter)	x		25:34:29
Turritella cooperi Carpenter	x	x	28-37:32
T. jewettii Carpenter	x	x	16-34:25

Locality 1: At an altitude of 700 feet on the ridge between Madranio and Javon canyons. Fossil-bearing sand is exposed in a road cut east of the wind gap crossing the ridge. Lat 34° 21′ N.

Locality 2: At an elevation of 500 feet on the county road to the summit of Rincon Mountain; N. 70° W. of Punta Gorda. Lat. 34° 22′ N.

* Molluscan ranges are from Keen (1937, p. 1-84). Last figure is the midpoint.

The fossils were collected by E. H. Quayle and identified by U. S. Grant, who wrote the following statement:

"The molluscan fauna collected from the Ventura marine terraces represents a shallow water association of somewhat mixed ecologic facies. Littoral and sublittoral species have been mixed, probably due to wave and current transportation. Wave action on dead shells is indicated by the abraded edges of some of the larger fragments of pelecypods. Some of the species now occur chiefly in the offshore zone in depths of five or ten fathoms or more, but, taken as a whole, the fauna indicates a shallow water environment.

"The living ranges of the species from these localities suggest that the temperature of the marine environment was but little different, if any, from that now prevailing at or near Ventura. The temperature may have been slightly lower but not sufficiently so to permit such cool water indicators as *Patinopecten caurinus, Chlamys jordani, Pandora grandis, Thyasira disjuncta, Thracia trapezoides* or *Mya truncata* to exist at this latitude as they did at San Pedro in Timms Point time. On the other hand, the southern forms present in the Palos Verdes faunas, such as *Dosinia ponderosa, Chione gnidia,* and *Laevicardium elatum* are also absent. Although there are some differences in faunal content between the present collections and those obtained from the Higgins Asphalt Pit locality (Grant and Strong, 1934, p. 1-5), the temperature significance of both are practically identical. All the species are living except *Cardita monilicosta* Gabb, which is very close to the living *Cardita nodulosa* (Dall), and since no decided change in temperature is indicated by the species present, the fauna suggests a post-Pleistocene age for the deposit."

CARPINTERIA TERRACE

General statement.—The fossil fauna and flora of the Carpinteria tar pit give valuable information concerning the climate at Ventura during the period of terrace cutting. The tar pit is a short distance west of

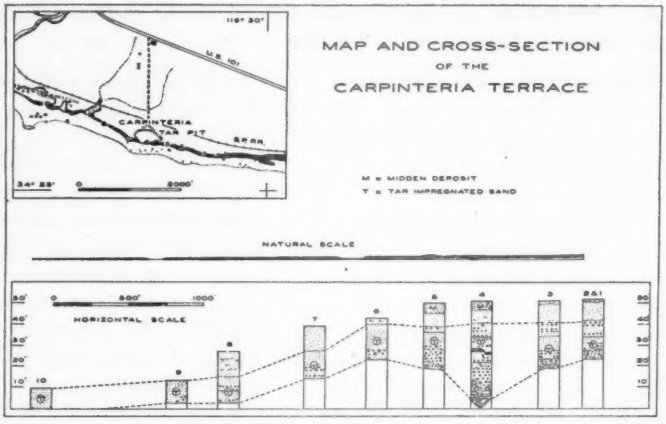

FIGURE 4.—*Map and sections of the Carpinteria terrace*

the area investigated most thoroughly and is at the seaward margin of the Carpinteria plain midway between the tracks of the Southern Pacific Railroad and the sea cliff about 1.5 miles west of Rincon Creek (Fig. 4, Pl. 5, fig. 2).

Brief mention of the discovery of fossils was made by Hoffman, *et al.* (1927, p. 155-156). The fossils and their inferred environment are described in detail in a series of recent articles: Grant and Strong (1934, p. 1-5) discuss the marine invertebrates; Miller (1931, p. 361-374) the vulturine birds; Miller (1932, p. 169-194) the passerine birds; Chaney and Mason (1933, p. 45-79) the plants; and Wilson (1934, p. 59-76) the land vertebrates. The consensus of these papers is that almost all the species are living, and that since the late Pleistocene there has been a decrease in humidity and an increase in temperature.

Paleontologic evidence.—The evidence of the Carpinteria fossils includes a number of contradictions, but most of these may be reconciled when the geologic setting is considered. The most important ecologic evidence is as follows: (1) The marine invertebrates indicate shallow, cool water, not markedly lower than present-day temperatures. (2) The passerine birds are typical of the region centering about the Mon-

terey Peninsula. (3) The larger birds are carnivorous, and their bones show little wear or decay. (4) The vertebrates include several whose habitat is semi-arid. The larger mammals are juvenile forms, and like the birds their bones show little attrition. (5) The plants grew in a more humid and cooler climate than the present. Most of the wood shows transportation, and this is substantiated by the number of unlike species, including seven trees, taken from a space 10 feet square.

Geologic setting.—The Carpinteria fossils are in tar-impregnated sand and gravel of the second marine terrace. This terrace is equivalent in age to the 200-foot terrace on Rincon Mountain 2 miles to the east. The cliff cut in the seaward margin of the terrace is 50 feet high at the asphalt quarry, and decreases westward until the terrace disappears below sea level at Carpinteria,—a decrease in elevation of 200 feet in 3 miles.

The terrace surface not only has been warped, but is also faulted, as shown in the road cut near the top of the grade west of the causeway crossing Rincon Creek. Here Miocene siliceous shale is thrust over unconsolidated dune sand (Kew, 1932, p. 66-67). The depression on the terrace surface between two ridges southwest of the highway on the strike of the fault is probably due to fault movement, as are low ridges and undrained depressions north of the coast highway west of Rincon Creek. The north-facing cliff 1 mile north of the highway is proved by wells to have a fault at its base.

The Carpinteria fossils, except the marine invertebrates, are in terrestrial gravels deposited by extended consequent streams that crossed the emerged sea floor. Gravels exposed in the cliff face thin westward—boulders and cobbles are more abundant near the east edge of the area; sand and silt towards the west. The cobbles are well rounded Eocene sandstone—none could be derived locally and their interpretation as products of marine abrasion is impossible. The sand lenses are cross-bedded and contain thin gravel layers. The tar-saturated layer in the middle and lower part is covered almost everywhere by fine sand and clay. Kitchen middens containing shells are on the surface of the Carpinteria terrace. East of section 6 (Fig. 4) the lower part of the terrace gravel contains little tar; west of section 6 most of the gravel and sand is saturated to the base.

The largest logs come from the vicinity of section 4. The logs are in tar-impregnated gravel immediately above the cobble and boulder fill of a small stream channel cut in Miocene shale underlying the Carpinteria terrace. This small stream channel after being excavated to present sea level, or slightly below, was completely filled with gravel to the level of the modern terrace surface. No trace of the former

course survives in the present topography. This history of channel cutting followed by filling preceded the retreat of the Carpinteria sea cliff to its present position. The alternation of degradation and aggradation may in part be related to Pleistocene fluctuations of sea level, but obviously is due in part to active local crustal movements. Detailed sections of the Carpinteria gravels are as follows:

	Feet
A measured section of the stream channel previously described (Fig. 4, section 4)	
(5) White sand and grayish-green clay overlain by light-gray to black silty sand containing shells in a midden deposit	12
(4) Cross-bedded, brown-weathering sand and lenses of sandy gravel	6
(3) Poorly sorted, tar-impregnated, cobble-gravel; cobbles up to 6 inches in diameter	3
(2) Cross-bedded sand, sandy silt, and gravel beds. Tar-impregnated layer containing logs is near base	8
(1) V-shaped channel filled with cobbles and boulders. The largest boulders are at base and are 3 to 4 feet in diameter	20
Total	49
A section 500 feet east of the asphalt quarry, typical of the Carpinteria terrace deposit (Fig. 4, section 5).	
(3) Bluish-gray to black, powdery sand and silt, grading to coarser brown sand at the base. Midden deposits cover the top	5
(2) White powdery sand with few pebbles	7
(1) Dark bluish-gray sand, prominently cross-bedded and with numerous cobble and gravel layers, particularly near the base. The upper third to half of this unit is tar-impregnated	15
Total	27

Environment of deposition.—The vertebrate remains seem to imply a different environment of deposition from the one inferred by Chaney and Mason from the plants. The fossil bird bones are minute and fragile, chiefly of carnivorous species for the nonpasserine forms; the larger vertebrates are principally juvenile forms. These fossils are considered by Miller and Wilson to be a normal tar pit accumulation. The plants could not have lived contemporaneously in the small area of the pit they were taken from, nor did they constitute an ecologic group at any station. The apparently conflicting evidence may perhaps be reconciled if present conditions in the Upper Ojai Valley are considered. Several active tar seeps close to Sisar Creek now trap animals as large as calves, as well as rodents and birds. Leaves are freely mixed with the tar, and during floods plants from higher in the Santa Ynez Mountains are transported into the tar-covered area and trapped.

The tar at Carpinteria enters the gravel from fractures in the underlying shale. The seeps were active in historic times, and were first described by Fr. Juan Crespi in 1769 (Bolton, 1927, p. 158-164):

"Just about north an arroyo comes down (Carpinteria Creek). My companion went to see it, and he says that it has a good stretch of water at the foot of the

mountains. The soldiers and the explorers said there was another good village of heathen. Not far from the town we saw some springs of pitch. The indians have many canoes, and at the time were building one, for which the soldiers named this town La Carpinteria, while I christened it with the name of San Roque . . ."

Sequence of events at Carpinteria.—A comparison of U. S. Coast and Geodetic Survey field sheets for 1869 and 1933 shows no significant change in the shore line. Hence it is impossible to estimate the rate of retreat of the sea cliff and the minimum age of the deposit in years. However, the relative position of the Carpinteria gravel in the Pleistocene sequence is determinable, and it may be possible some day to correlate the Carpinteria brea deposit with similar traps at McKittrick and Rancho la Brea.

In the following sequence, the events are listed from oldest to youngest:

(1) Deposition of 500 feet of sandy silt and sand in the late Pliocene or early Pleistocene following a period of deformation in the early Pliocene or late Miocene.

(2) Deformation in a second orogeny in the medial Pleistocene, characterized in this area by the development of south-dipping, high-angle reverse faults.

(3) Stream erosion which continued long enough to carve a surface of late maturity in areas of weaker rocks.

(4) Intermittent uplift and cutting of nine marine terraces on Rincon Mountain. The Carpinteria terrace is next to lowest.

(5) Withdrawal of the sea to beyond the present shore line.

(6) Deposition of stream gravel and its impregnation with tar.

(7) Cutting of the modern sea cliff and entrenching of streams crossing the terrace surface.

Deformation has been active throughout this time and continued into the Recent. The mingling of arid-climate species as *Dipodymys* with plants resembling the Monterey Peninsula flora may be explained by the climatic amelioration of postglacial time. Another important factor was the introduction by streams of plants from higher stations of cooler temperature in the Santa Ynez Mountains.

STRUCTURE

GENERAL CONSIDERATIONS

The principal structures of the Ventura region are listed below according to the physiographic province in which they occur:

Subdivision	Major structures	Rocks involved
Santa Ynez Mountains	Matilija overturn	Eocene sandstones
	San Cayetano thrust	Eocene thrust over Miocene and Pliocene
Ojai Lowland	Ojai syncline, Lion Mountain anticline, Santa Ana Valley folds, Santa Ana fault	Oligocene (?) and Lower Miocene rocks

Subdivision	Major structures	Rocks involved
Sulphur Mountain Upland	Thrust faults and complex minor folds	Miocene shales
	Red Mountain dome	Oligocene (?) rocks
	Red Mountain thrust	Oligocene (?) thrust over Miocene and Pliocene
Coastal Hills	Ventura Avenue anticline	Pliocene and Pleistocene strata
	Canada Larga syncline	
	Wheeler Canyon homocline	

SANTA YNEZ MOUNTAINS

Matilija overturn.—The Matilija overturn, the principal structure on the southern slope of the Santa Ynez Mountains, is the overturned south limb of a large anticline (Kerr and Schenck, 1928, p. 1087-1102). The rocks involved are the undifferentiated Eocene and Matilija, Cozy Dell, and Coldwater formations. They appear in inverted order above the Sespe formation exposed at the base of the range. The sequence is normal west of Santa Ana Creek and the strata dip 40° to 45° S. East of Sisar Creek the structure is broken on the south limb by the San Cayetano thrust. Overturning ends east of this creek and the fold is a broad, eastward-plunging anticline.

San Cayetano thrust.—The San Cayetano thrust (Kew, 1924, p. 100-101) entering the Ventura region east of Timber Canyon and disappearing to the west in the alluvium of the Ojai Valley has been traced 20 miles eastward. It is a high-angle thrust fault with the north side upthrown. South of Santa Paula Ridge the fault dips 45° N., and Matilija sandstone is thrust over "Santa Margarita" and Pico shale.

The San Cayetano thrust is strongly expressed in the topography east of Santa Paula Creek where rocks brought in contact by the fault differ greatly in erosional resistance. The Matilija sandstone of the overthrust plate makes a 2000-foot composite fault-line scarp above the weaker "Santa Margarita" and Pico shale at the base of Santa Paula Ridge. West of Sisar Creek the topographic effect of the fault is less pronounced as rocks brought in contact are of approximately equal resistance.

The San Cayetano thrust is probably still active. Terrace gravels north of the fault east of Sisar Creek are tilted 25° N. towards their source in the Santa Ynez Mountains. It is possible that the difference in elevation of 250 to 300 feet between the Upper Ojai Valley and the dissected fans on either side of Sisar Creek is the result of uplift of the fans by the fault. The evidence is largely topographic, but the fans are truncated at the fault, and streams crossing the fans are entrenched in valleys 250 to 300 feet deep.

OJAI LOWLAND

General statement.—The structure of the Ojai Valley appears to be synclinal, but the structural details are obscured by terrace and alluvial fan gravels. The Lion Mountain anticline is on the southern margin of this valley, and east of this fold underlying the ridge separating the Ojai and Upper Ojai valleys is an overturned and faulted syncline (Pl. 1). The structure of the Santa Ana Valley, the western extension of the Ojai Valley, differs from the main valley and will be discussed first.

Santa Ana Valley structures.—The dominant structure of the western Santa Ana Valley is a south-dipping homocline in the Sespe formation, dipping 20° to 30° on the average, and reaching a maximum of 65° at the base of the Santa Ynez Mountains.

This relatively simple structure is interrupted by a number of minor folds and two important faults (Pl. 1, section A-A′). The south side is upthrown on both of them. The southern of the two faults, the Santa Ana, extends east 4.5 miles from the junction of Coyote Creek and its west fork, and is responsible for the abrupt reversal of dip from 20° S. to 70° N. on the north side of the east-trending cuesta that divides the Santa Ana Valley through the center. Terrace gravel east of the Ventura River obscures the eastward continuation of the fault. Probably the fault is at the base of the gravel-covered ridge south of Long Valley, and continues eastward north of the dividing ridge between the two Ojai valleys (Pl. 1, section B-B′).

If this assumption is correct, the Santa Ana fault cuts off the north limb of the Lion Mountain anticline. The direction of displacement, —the south side is upthrown—is opposite the San Cayetano, and the Santa Ana fault is overridden by it. The 250- to 300-foot difference in elevation between the Upper Ojai and Ojai valleys may be the result of erosion by San Antonio Creek of a fault-line scarp along the Santa Ana fault.

Three folds are south of the Santa Ana fault, from north to south: the Santa Ana syncline, the Coyote Creek anticline, and the Ayers Creek syncline (Pl. 1). The moderate dips on the flanks of these folds average 30°, and each fold is approximately 4 to 5 miles long. The axis of the Ayers Creek syncline is followed by a narrow inlier of Miocene rocks bordered by the Sespe formation. These folds plunge east and cross the Ventura River, although their extension through the faulted area beneath the gravel-covered Oakview terraces is uncertain.

Ojai Valley structures.—The structure of the Ojai Valley is masked by terrace gravel, alluvial fans, and stream deposits. The valley is

bordered on the north by the Matilija overturn and on the east by the San Cayetano fault. On the south the valley is limited by the overturned Lion Mountain anticline and by the Santa Ana fault. These relationships are shown by structure section B-B′ (Pl. 1), and resemble the structure of the Santa Clara Valley north of Oak Ridge (Reed and Hollister, 1936, p. 101), where marginal thrust faults dip in opposite directions away from the valley.

Lion Mountain anticline.—The Lion Mountain anticline exposes the Sespe formation for 2.5 miles in the hill that separates Lion Canyon from the Ojai Valley. The anticline plunges east, as do most of the folds in the Ventura region, and is overturned north. The Vaqueros borders the Sespe on the south and east sides but is cut out on the north and northwest flanks by the Santa Ana fault.

Upper Ojai Valley structures.—As in the Ojai Valley, the structure of the Upper Ojai Valley is obscured by terrace gravel. The low ridge separating the valleys is a northward-overturned syncline that exposes lower Miocene and Oligocene (?) rocks; the Modelo forms the core. Faults cut the fold and reduce the outcrop width of formations on the flanks, particularly on the east limb where the Sespe and Rincon formations crop out as narrow bands. This syncline, if projected across the Upper Ojai Valley, should appear on the north slope of Sulphur Mountain, but it is not found there. It is probably truncated by faults beneath the valley floor.

The south-dipping Sisar fault on the north slope of Sulphur Mountain and in the canyon of Sisar Creek brings the Modelo formation in contact with the "Santa Margarita." This is well shown at the junction of the east fork of Santa Paula Creek and Sisar Creek. The north-dipping San Cayetano thrust cuts and overrides the Sisar and other south-dipping reverse faults (Pl. 1, section C-C′). The San Cayetano probably has the same relation to the Santa Ana fault, although this intersection is obscured by alluvium. Thus there is evidence of two periods of faulting during the medial Pleistocene diastrophism—with the south-dipping reverse faults preceding the north-dipping San Cayetano.

SULPHUR MOUNTAIN UPLAND

General statement.—As a primary result of the incompetence of the Miocene siliceous shales—the principal rock type cropping out on Sulphur and Rincon mountains—the most complex structures are in this province. Sulphur Mountain is essentially a south-dipping homocline near the center, but this structure is complicated by many faults and

minor folds. Red Mountain dome, west of the Ventura River, is an anticline with the Sespe formation exposed in the core. The south flank is overturned and is broken by the Red Mountain thrust, a major structure of the province.

Sulphur Mountain structures.—The structure of the central part of the mountain is a south-dipping homocline of the four Miocene formations: the Vaqueros and Rincon on the north side, the Modelo on the central ridge, and the "Santa Margarita" on the south. East of the western third of the mountain the south limb of the structure is overturned and the strata dip north in inverted sequence (Pl. 1, sections B-B′, C-C′).

The structure is complex at the west end of the mountain where shallow folds of the southern Santa Ana Valley cross the Ventura River and extend into the mountain. The Ayers Creek syncline crosses the river at the mouth of San Antonio Creek and its core of "Santa Margarita" formation extends westward across the mountain at the head of Fresno Canyon. In the southern extension of Sulphur Mountain paralleling the Ventura River the strike swings from east-west to north-south owing to the eastward plunge of the Red Mountain dome.

Although the bold escarpment on the south side of Sulphur Mountain has been interpreted as a fault scarp (Kew, 1932, p. 61-62), there is no evidence of faulting at the base of the cliff. At the head of Wheeler Canyon and eastward to Santa Paula Creek the contacts between the "Santa Margarita" and Pico formations are gradational, and any fault of sufficient magnitude to account for the scarp can be definitely excluded. The "fault scarp" is best interpreted as an erosional escarpment due to rapid erosion of the weak "Santa Margarita" shale as compared to the resistant Modelo shale. Locally there has been some faulting, particularly in Hammond and Aliso canyons where the Modelo formation is thrust over the "Santa Margarita," but this merely reduces the outcrop width of the "Santa Margarita."

The eastern half of Sulphur Mountain is an east-plunging anticline with the north limb dropped by the Sisar fault. Owing to the eastward plunge, the Pico and "Santa Margarita" formations swing around the eastern end of the mountain at Santa Paula Creek and extend westward on the north side to the Upper Ojai Valley. The Pico formation also crops out in Big Canyon where it has been faulted down. Headward erosion along these weak rocks has localized Santa Paula Creek in its present course. Similarly at the west end of the mountain the Ventura River originally crossed the Sulphur Mountain Upland in the north-striking Rincon shale at the east end of the Red Mountain dome.

Red Mountain dome.—The Red Mountain dome is a doubly-plunging anticline with moderately resistant shales and sandstones of the Sespe formation exposed in the core. The Vaqueros oyster reef nearly encircles the dome, except for a few places on the south side of the fold where it is eliminated by faulting. The dome deflects the outcrop of the Miocene strata on the south limb of the fold 2.5 miles south of the outcrop of the same formations on Sulphur Mountain.

The north limb of the fold is complicated by two minor folds,—a syncline and an anticline whose axes parallel the main structure. The south limb is overturned, and in deep canyons the Sespe formation may be seen to dip north and overlie the Vaqueros and Rincon formations. The overturning is related to the Red Mountain thrust. This fault overrides the Canada Larga syncline west of the Ventura River, and brings the south limb of the Red Mountain anticline in contact with the north limb of the Ventura Avenue anticline (Pl. 1, section A-A′). The dome pitches eastward between 35° and 45°, and no Sespe strata are exposed east of the Ventura River, although they form prominent cliffs on the west bank.

Red Mountain thrust.—The Red Mountain thrust extends westward on the south side of Red and Rincon mountains for 13 miles from the western end of Sulphur Mountain to the sea at Rincon Point. The fault trace curves sharply from north-south east of the Ventura River to east-west on the west side of the river. This change in trend is closely related to the Red Mountain dome (Pl. 1). The fault dips north or west at 40° to 65° through most of its length and thrusts Miocene strata over Pliocene, except near Padre Juan Canyon where lowermost Pico ("Repetto") shales are brought in contact with uppermost Pico clay shales.

The thickness of each Miocene formation cropping out on Sulphur Mountain averages about 2000 feet. West of the Ventura River the same formations north of the fault are attenuated fault-slivers only a few hundred feet thick. There is little breccia along the fault, despite the displacement of several thousand feet, and evidence of faulting is often difficult to find. This is true in Padre Juan Canyon where Pliocene shales, lithologically nearly identical and normally several thousand feet apart, are in fault contact. The fault is marked by only a 6-inch band of gouge exposed in an excavation for an oil well at the head of a small tributary of Madranio Canyon. Here the sharply defined fault plane dips 65° N.

The Red Mountain thrust is still active and has displaced the 500-foot marine terrace on Rincon Mountain and arched the Ventura River

terraces. The 800- to 900-foot river terrace south of Fresno Canyon is at the north end of the thrust and is tilted north opposite the original stream flow (Pl. 3, fig. 3).

COASTAL HILLS

General considerations.—Rocks underlying the Coastal Hills are poorly indurated shale, sandstone, and conglomerate of the Pico and "Saugus" formations. These weak rocks were deformed into broad structures compared with the complex folds and faults of Sulphur Mountain. The three major structures of the Coastal Hills are: the Canada Larga syncline, the Ventura Avenue anticline, and the Wheeler Canyon homocline.

Canada Larga syncline.—The axis of the Canada Larga syncline, 4.5 miles north of Ventura, extends east 7.5 miles from the Ventura River to Wheeler Canyon where the fold disappears under the alluvium of the Santa Clara Valley. Dips on either limb of the syncline range from 30° to 50°. The fold plunges east and at the eastern end a tongue of "Saugus" is folded in along the axis.

The syncline ends at the Ventura River where the fold is overridden by the Red Mountain thrust. That the syncline axis probably continues beneath the fault is indicated by the contact between the upper and lower Pico which crosses the Ventura River without significant offset south of the fault. At Weldon Canyon upper Pico shales strike east-west and pass under the fault without appreciable drag.

Ventura Avenue anticline.—The Ventura Avenue anticline is traceable from Wheeler Canyon west for 15 miles until it disappears in the Pacific Ocean. The general trend of the axis is east-west, although there is a southwardly convex curve where the fold crosses the Ventura River. West of Canada del Diablo the axis trends N. 67° W. at a slight angle to the coast. The anticline plunges east at 15° to 20°; east of the Ventura River the axial plane dips south, but west of the river the dip is north where the fold is influenced by the Red Mountain thrust.

The north limb of the anticline is faulted at Padre Juan Canyon by a south-dipping reverse fault that is overridden at the eastern end by the Red Mountain thrust. The reverse fault brings middle Pico sandstone on the hanging wall in contact with upper Pico clay shale on the northern side of the fault. On the northern side of the downdropped block the Red Mountain thrust brings upper Pico shale in contact with lower Pliocene ("Repetto") shale. These two reverse faults on the north limb of the anticline cut out the thick section of middle Pico sandstones and

FIGURE 1. MARINE AND RIVER TERRACES AT VENTURA (SPENCE AIR PHOTOS)

FIGURE 2. VENTURA RIVER TERRACES SOUTH OF LONG VALLEY (SPENCE AIR PHOTOS)

FIGURE 1. OAKVIEW TERRACE
At junction of Ventura River and San Antonio Creek (Spence Air Photos).

FIGURE 2. TERRACES WEST OF VENTURA RIVER MOUTH (SPENCE AIR PHOTOS)

FIGURE 3. TERRACES OF UPPER OJAI STREAM SOUTH OF FRESNO CANYON

substitute for them the weak shales of the uppermost and lowermost Pliocene.

There is strong topographic contrast between the upper Pico clay shale and the middle Pico sandstone, but none between the upper and lower Pico shales. Hence, the prominent sandstone hogbacks on the seaward flank of the anticline at Pitas Point are missing on the north limb, and in their place are rounded hills underlain by Pico shale.

The Ventura Avenue anticline is broken by numerous minor faults. Reversals in dip and contorted strata are well displayed north of Canada del Diablo. Several small faults may be recognized in the hills behind Ventura. Wells drilled in the Ventura Avenue field prove that numerous faults exist, some of large displacement.

Eaton (1929, p. 713-762) and Hertel (1928, p. 721-742) directed attention to the great thickness of Pliocene and Pleistocene strata involved in the fold. Wells drilled to 10,000 feet are in rocks no older than Pliocene. Dips on both limbs of the structure range from 20° to vertical, with the average probably between 35° and 40°. The axial band of horizontal strata is narrow, as beds change from a south dip of 40° to a north dip of 45° in 500 to 700 feet. This sharp flexure is shown in Figure 2 of Plate 4.

Throughout most of its length transverse ridges crossing the anticline are lower where they cross the axis than on the flanks. Several subsequent streams, notably those tributary to Canada del Diablo and Hall Canyon, parallel the axis. The curvature of the conglomerate and sandstone ridges between Peppertree and Hall canyons reflects the eastward plunge of the anticline. West of Aliso Canyon the outcrop of the "Saugus" formation and the southern limit of the Coastal Hills is shifted nearly 2 miles south of the corresponding positions in the homoclinal section east of the canyon.

Wheeler Canyon homocline.—A south-dipping homocline of Pleistocene, Pliocene, and upper Miocene strata underlies the Coastal Hills between Wheeler Canyon and Timber Canyon (Pl. 1, section C-C′). In the southern part of the homocline dips are south and range from 30° to 89°. The dips increase northward from the Santa Clara River, until at the southern base of Sulphur Mountain the strata pass from vertical to overturned and dip north.

The principal complications in this essentially uniform structure are sharp reversals of dip, probably due to faulting, north and east of Canada Lārga. The line of overturning is farther south in the Coastal Hills east of Santa Paul Creek and is approximately at the midpoint of the section, doubtless owing to the San Cayetano fault.

SULPHUR MOUNTAIN SURFACE

GENERAL STATEMENT

A late mature subaerial erosion surface bevels the structures of the Ventura region. It was cut about the middle of the Pleistocene following the principal orogeny. This erosion surface is best preserved on the summit of Sulphur Mountain where it truncates the steeply dipping shales of the Modelo formation. Because of this, Sulphur Mountain surface appears to be more satisfactory than Timber Canyon surface, the name proposed by Gale (1931, p. 38):

"All these occurrences taken together seem to outline in a general way an old land surface that has been subjected to considerable erosion since the deformation of the region and that was uplifted essentially undisturbed and cut into by streams with a lower base level. For this surface the name Timber Canyon surface is proposed because one of its most clearly preserved remnants is the Timber Canyon fanglomerate."

On page 37 he states that:

"This old surface is marked by high-level terraces and deposits of older alluvium or fanglomerate that were formed at the basis of the upthrust mountains. One of the best of these remnants is a deposit of reddish fanglomerate occurring on the tops of ridges and bevelling the edges of the upturned marine beds on the north side of the Santa Clara Valley. This deposit was formed at the base of Santa Paula Peak, a mountain composed of abnormally hard Eocene strata with a present altitude of almost 5000 feet, and was folded and upthrust during the mid-Pleistocene diastrophism. As this fanglomerate surface passes undisturbed across the trace of the fault, it is evident that the movement along the fault had ceased before the deposition of the fanglomerate. . . ."

A major objection to the term, Timber Canyon surface, is that the Timber Canyon fan (3 miles east of Santa Paula Creek) is not a surface of erosion, but of deposition. The surface to be described was produced by stream erosion, and if Gale's term is used, it should be restricted to the later deposit made on the surface. The two are not of equivalent age,—the Timber Canyon fanglomerate fills a valley incised below the "Timber Canyon surface" after the carving of the surface. Similar valleys have been excavated in the Coastal Hills east of Santa Paula Creek since the cutting of the surface and later filled with gravels derived from Santa Paula Ridge. At least four stages of valley excavation and filling are represented, and these stages may correlate with pulsatory uplift that followed the cutting of the Sulphur Mountain surface.

A minor objection to Gale's statement is that Recent faulting has occurred along the San Cayetano thrust. It is true that the Timber Canyon fan was deposited long after the principal movement on the thrust, but the tilting and faulting of similar gravels near Sisar Creek disproves the inactivity implied by Gale.

FIGURE 1. VENTURA FLOOD PLAIN AND DELTA (FAIRCHILD AERIAL SURVEYS, INC.)

FIGURE 2. ANTICLINAL AXIS 3 MILES WEST OF VENTURA RIVER (SPENCE AIR PHOTOS)

VENTURA RIVER AND VENTURA AVENUE ANTICLINE

FIGURE 1. MARINE TERRACES ON RINCON MOUNTAIN (SPENCE AIR PHOTOS)

FIGURE 2. CARPINTERIA ASPHALT PIT AND TERRACES (SPENCE AIR PHOTOS)

COASTAL PHYSIOGRAPHY

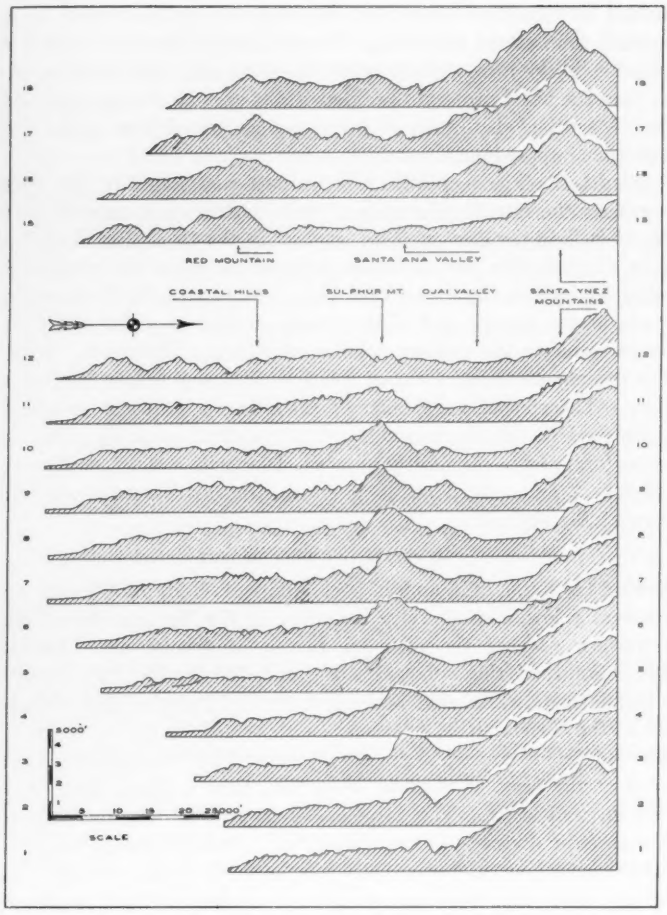

FIGURE 5.—*Projected profiles*

PROJECTED PROFILES

The projected profiles of Figure 5 illustrate the characteristics of the Sulphur Mountain surface. As shown on the profiles, the surface had a relief of 2500 to 3500 feet, with the highest elevations in the Santa Ynez Mountains.

The profiles are on north-south lines spaced 1 mile apart and consist of the highest ridge crest in each 1 mile area projected on the eastern

boundary of the strip. They are essentially a sky-line profile for each mile-wide zone across the area. This method yields more satisfactory results for the Ventura district, with its strong relief and complex structure, than the usual projected profile with its greatly exaggerated vertical scale. A vertical distortion of twice the horizontal was used to avoid this artificial appearance.

Profiles 1-12 (Fig. 5, profile 1 is farthest east), east of the Ventura River, show the superior elevation of Sulphur Mountain over the Coastal Hills, as well as the erosion scarp separating the two. The Ojai Valley, largely excavated in the Oakview epicycle, is shown on profiles 8-12. Profiles 3-7 include the Upper Ojai Valley, and on profile 2 the extensive terraces at the eastern end of the valley appear as a flat bench at an altitude equal to the eastern summit of Sulphur Mountain. Profile 8 shows the asymmetrical form of the low ridge separating the two Ojai valleys. The steeper north side of the ridge is probably a resequent fault-line scarp. Hogbacks made by the resistant members of the upper Eocene formations are indicated near the north end of profiles 6-16; profiles 6-12 cross the Matilija overturn and 15-18 are in the normal section of the anticline.

DESCRIPTION OF THE SURFACE

During the Sulphur Mountain cycle, Sulphur, Red, and Rincon mountains were a low, sinuous ridge breached by the Ventura River, and in the western part of the area by Javon, Madranio, and Los Sauces creeks. South of this subdued ridge, and 500 to 700 feet lower, was the undulating lowland that beveled the folded Pliocene and Pleistocene rocks of the Coastal Hills.

The Sulphur Mountain segment was crossed by streams whose valleys headed at the crest of Sulphur Mountain, then a short distance north of the present crest. The principal exception to the more northerly position of the divide is the summit ridge of Sulphur Mountain between Sulphur and Aliso canyons. This part of the original divide is about half way along the ridge, and is convex southward as a result of the doming of the Lion Mountain anticline. South-flowing streams elsewhere have been truncated by the more rapid extension of the north-flowing streams draining into the Upper Ojai Valley and Lion Canyon.

The Ventura River was established in about its present position during the Sulphur Mountain cycle, and its tributary streams excavated the Ojai Lowland to within 300 to 500 feet of the present depth. Sisar Creek and the east fork section of Santa Paula Creek crossed the Upper Ojai Valley to join the Ventura River. This west-flowing stream was not captured by the Coastal Hills portion of Santa Paula Creek until the

close of the Sulphur Mountain cycle. The details of this capture are described in the section on physiography of the Upper Ojai Valley.

The Coastal Hills provide little direct evidence of the Sulphur Mountain surface. Almost all the surface has been converted into slopes, and although the ridges are nearly accordant, they have been lowered several hundred feet below their probable altitude at the close of the late Pleistocene regional uplift. The chief exceptions are the boulder-covered ridges surrounding the 1950-foot peak between Canada Larga and Aliso Canyon. The boulders rest on the truncated edges of upper Pico conglomerate, and were deposited on the Sulphur Mountain surface by streams with a regimen not greatly different from the present ones. The boulders were supplied by erosion of the conglomerate.

PHYSIOGRAPHIC PROVINCES OF THE VENTURA REGION

GENERAL CONSIDERATIONS

The drainage pattern is partly inherited from the Sulphur Mountain cycle and is partly the result of stream adjustment in the present cycle. Most of the adjustment was accomplished by headward erosion and capture by south-flowing streams of those with gentler gradients that flowed east or west. The result is the development of a few large through-flowing streams whose courses are normal to the structural trends of the region.

SANTA YNEZ MOUNTAINS

All the streams of the southern Santa Ynez Mountains, except the Ventura River, are restricted to the southern slope, and are short with steep gradients. They flow at right angles to the strike of the Eocene formations and are tributary to either the Ventura River or Santa Paula Creek.

Subsequent tributaries to the main streams are best developed in the Cozy Dell shale and have excavated a nearly continuous lowland paralleling the front of the Santa Ynez Mountains. This lowland, between the resistant Coldwater and Matilija sandstones, is most prominent in the overturned sections north of the Ojai Valley and also west of the Ventura River in Kennedy Canyon.

The overturned Coldwater sandstones make a barrier on the south side of the Cozy Dell lowland east of the Ventura River. Small streams flowing south are deflected at the Coldwater contact and flow east or west until they join one of the larger through-flowing streams. This Coldwater barrier has decreased the gradient of the small streams and compelled them to construct large alluvial fans against the north face of the hogback.

OJAI LOWLAND

General statement.—The drainage of the Ojai Lowland has evolved through excavation of the Ojai Valley by west-flowing tributaries of the Ventura River and their capture by the headward growth of south-flowing streams. These streams are likewise tributary to the Ventura River, but join it south of the Ojai Valley at a lower elevation. In addition to the drainage changes, the interpretation of the Ojai Lowland is complicated by warping and faulting, both of which are still active.

The Ojai Lowland, including the Santa Ana Valley as part of the same basin, may be divided into three sections. The Santa Ana Valley is essentially a stripped surface. Bedrock is exposed over wide areas, or is covered with residual soil. Bedrock in the central, or Long Valley, section of the Ojai Valley is concealed by river terraces. The eastern Ojai valley is filled with alluvium, and at the eastern end are the large alluvial fans of Señor and Horn canyons.

Santa Ana Valley.—The Santa Ana Valley is crossed by Santa Ana and Coyote creeks and is divided into two nearly equal sections by an east-trending cuesta with the Santa Ana fault on the north. Santa Ana Creek crosses the ridge through a water gap 0.3 mile wide, while Coyote Creek crosses it in a narrow, steep-walled gorge whose rectilinear pattern is controlled by the strike of the Sespe formation. The longer, nearly east-west reaches of the gorge are cut in shale; the sandstone beds are crossed normal to the strike.

Santa Ana Creek was probably established by headward erosion during the Sulphur Mountain cycle, but of this history no trace survives today. The stream excavated its present valley in the period of uplift that started the present erosion cycle and cut a broad valley during the Oakview epicycle. The stream incised a narrow arroyo 50 feet deep as a result of the most recent uplift.

The south end of the Santa Ana Valley is covered with gravel deposited by Santa Ana Creek during the broad valley stage. This gravel bench is equivalent to the Oakview strath of the Ventura River. The Santa Ana terrace is tilted north in similar fashion to the Ventura terraces and decreases in altitude from 600 feet at the junction of Coyote and Santa Ana creeks to 500 feet at the south end of the Santa Ana Valley, a gradient of 1.6 per cent.

Coyote Creek was established later than Santa Ana Creek by headward erosion and diversion of east-flowing tributaries of Santa Ana Creek. The first to be diverted in the lower Santa Ana Valley was captured by extension of a tributary of Willow Creek through the ridge of

Sespe sandstone at the southern boundary of the Santa Ana Valley (Fig. 6). The evidence is the gravel-covered floor of the abandoned stream valley between the elbow of capture and Santa Ana Creek. This wind gap is the low divide at 500 feet between Coyote and Santa Ana creeks in the western half of the southern Santa Ana Valley. The capture occurred very shortly after the Oakview epicycle, as shown by the stream terrace at an elevation of 450 to 500 feet on the east side of Coyote Creek at the confluence with Santa Ana Creek. This terrace, common to both streams, is a correlative of the first terrace below the Oakview strath of the Ventura River, and was cut by Coyote Creek after its present course was established. The stages in this history of drainage evolution are illustrated in the diagrams of Figure 6.

Streams tributary to Santa Ana Creek in the northern Santa Ana Valley were captured by Coyote Creek in similar fashion. The headward extension of a tributary of Coyote Creek through the medial ridge tapped the largest western tributary of Santa Ana Creek. The surviving trace of the beheaded tributary is the gravel-covered valley that extends westward 1.4 miles to Coyote Creek from Santa Ana Creek.

The last event in the drainage history of the northern Santa Ana Valley was the capture of Poplin Creek, a small stream that originally flowed southeast to join Santa Ana Creek. Poplin Creek now flows southwest to Coyote Creek through a narrow gorge 0.8 mile east of where Coyote Creek enters the Santa Ana Valley.

The drainage history of the Santa Ana Valley east of Santa Ana Creek is indeterminate. Both halves of the valley open eastward into the Ventura River valley 100 feet higher than the present flood plain. Both these broad valleys are occupied by underfit streams; the northern drains eastward into the Ventura River, the southern flows west to Santa Ana Creek.

The two halves of Santa Ana Valley east of Santa Ana Creek may have been partially excavated by tributaries of the Ventura River when it was at the level of the 1000-foot terraces. These tributaries possibly were captured near the close of the Sulphur Mountain cycle by headward growth of Santa Ana Creek. Any gravels deposited by these east-flowing streams in the Sulphur Mountain cycle have been stripped away.

Long Valley.—The gravel-covered terraces in the triangular area between San Antonio Creek and the Ventura River are the Long Valley section of the Ojai Lowland—so named for the valley followed by the Southern Pacific Railroad to Ojai.

Older dissected fans are at the base of the Santa Ynez Mountains north of Ojai. The largest is between Stewart and McDonald canyons

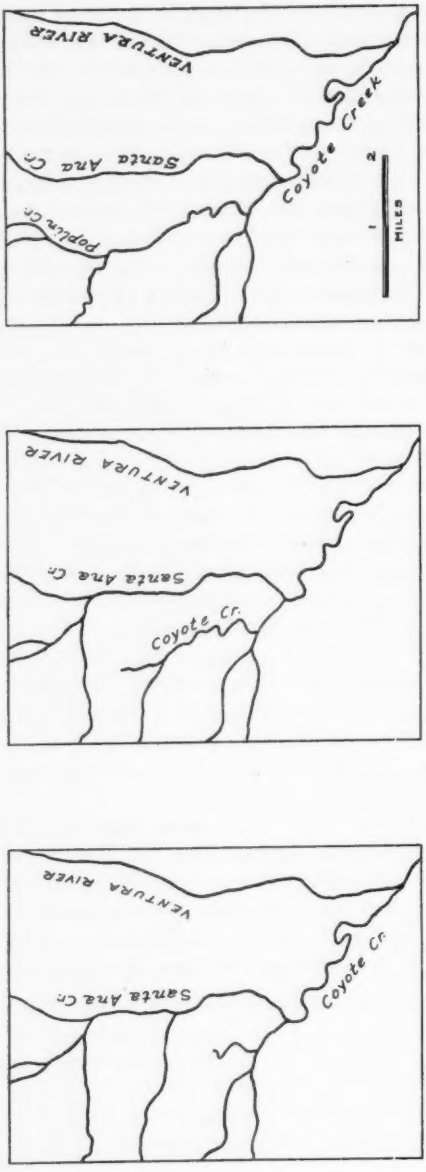

FIGURE 6.—*Drainage evolution of the Santa Ana Valley*

(Fig. 2) and slopes from an elevation of 1200 feet down to 950 feet. The Sespe formation is exposed beneath the gravel in canyons incised 200 feet below the surface of the fan. This inactive fan is a correlative of the higher Ventura terraces at Oakview.

East of this high-level fan is the more recent abandoned fan of Stewart Creek that slopes from 1350 feet to 800 feet at Ojai. Stewart Creek following the east margin of the fan has incised a narrow ravine southeastward across the apex. The capture of the surviving drainage of the fan, and eventually of Stewart Creek, by headward erosion of a small tributary of McDonald Creek appears imminent.

The toe of the Stewart Canyon fan, together with the Long Valley terraces is a barrier that divides the Ojai Valley into two parts. West of this north-south watershed the runoff is to the Ventura River, east of it the drainage of the Ojai Valley is by San Antonio Creek. The lowest elevations on the divide are at the east end of Long Valley (700 feet) and the parallel valley to the north (750 feet). These valleys are abandoned channels of west-flowing streams (Fig. 9), once tributary to the Ventura River, but now tributary to San Antonio Creek. Separating the two valleys is a low gravel-covered ridge, known locally as Krotona Hill (Fig. 2), that trends northwest 1.2 miles and is 150 feet high. Bedrock is hidden by the reddish terrace gravel, but is probably the Sespe formation. The asymmetrical profile, with the steep face north, together with the constriction of the Ventura River flood plain immediately west suggest that Krotona Hill is on the upthrown side of the continuation of a fault south of the Santa Ynez Mountains west of the river (Pl. 1).

Ventura River terraces in the triangular area between the river and San Antonio Creek are tilted northeast toward the eastern Ojai Valley. The origin of these terraces and the drainage evolution of San Antonio Creek are discussed in the section on Ventura River terraces.

Eastern Ojai Valley.—No terrace gravels are in the Ojai Valley east of a north-south line half a mile east of Ojai. The conspicuous feature in this part of the valley is the large alluvial fans of Señor and Horn canyons at the east end. These fans are covered with gray unweathered sandstone boulders in contrast to the oxidized soil and weathered boulders of the inactive fans on the north-central border of the valley.

The following points deserve emphasis in a discussion of the origin of the Ojai and Santa Ana valleys:

(1) The floor of the Santa Ana Valley is mostly bedrock. The thin soil is residual and is chiefly derived from the Sespe formation.

(2) The Ventura River is entrenched 50 to 100 feet below the Ojai

Valley and crosses it at right angles. Very little runoff from the Ojai Valley reaches the river directly.

(3) Stream terraces and dissected alluvial fans are in the central part of the lowland. These are higher than the eastern valley floor and the terraces are tilted north-east against the gradient of the Ventura River.

(4) The Ojai Valley had a dual outlet by Long Valley and the parallel valley north of Krotona Hill. The capture of the east end of the Ojai Valley by San Antonio Creek destroyed this connection with the Ventura River.

(5) The dissection of the fans on the north side of the Ojai Valley decreases eastward.

(6) The structure of the east end of the Ojai Valley is probably synclinal. This is south of the maximum overturning of the Matilija overturn and on the strike of the San Cayetano thrust.

This evidence indicates that the eastern half of the valley is downwarped and is an area of aggradation. The down-warping is probably related to continuing deformation of the Matilija overturn and the San Cayetano thrust. This activity is expressed topographically by growing alluvial fans in the eastern end of the valley, by drainage reversal at the west end, and by back-tilting of river terraces at the southern margin of the valley.

Upper Ojai Valley.—The Upper Ojai Valley is a lens-shaped basin 5 miles long and 1 mile wide; its altitude ranges from 1250 feet at the west end to 1550 at the east. The Upper Ojai Valley is separated from the Ojai Valley to the north by an asymmetric ridge, 300 to 500 feet high on the north and 50 to 150 feet high on the south. The ridge increases in height westward to Lion Mountain where the maximum relief is 1035 feet.

The difference in elevation between the two valleys probably results from erosion by San Antonio Creek along the eastward extension of the Santa Ana fault at the northern base of the dividing ridge. The south side of the fault is upthrown and brings comparatively resistant Sespe sandstone into contact with weaker Miocene shales. Tilted terrace gravels on the fault line south of Ojai show that renewed activity has occurred on this fault in the late Pleistocene.

Lion Creek is cutting headward in a rock-walled gorge at the west end of the valley, and follows the strike of a shale bed in the Sespe formation. This canyon by incising itself across the valley floor is lowering the Upper Ojai Valley from the present elevation of 1250 feet to the temporary local base level of San Antonio Creek at 550 feet. In follow-

ing the strike of the Sespe shale the stream has shifted its course approximately 1000 feet north of its position at the time the headwaters were captured by Santa Paula Creek. Lion Creek is separated by a strike ridge of Sespe sandstone from the gravel-covered wind gap that marks the former channel at an altitude of 1270 feet.

Stream terraces extend from the Upper Ojai Valley to the east fork of Santa Paula Creek, and are best preserved on the north side of the valley (Pl. 1). East of Sisar Creek the terraces are at higher elevations than corresponding terraces in the Upper Ojai Valley, as they have been uplifted by later displacement on the San Cayetano thrust—this displacement is responsible for the truncation of older alluvial fans at the mouth of Sisar Canyon.

A prominent terrace covering about 3 square miles is east of the fault between Bear and La Broche canyons (Pl. 1) at an average elevation of 2000 feet. The northern margin of the terrace is buried by alluvial fans built on it by streams from the Santa Ynez Mountains. These fans and the terraces are now deeply dissected by the streams that once built them, but that are now graded to Sisar Creek. The gravels have been faulted, and in one of the deeper canyons dip north 25° in a direction opposite to the supply.

The fact that terraces extend from the Santa Paula Creek westward across the Upper Ojai Valley, and in Lion Canyon to the Ventura River terraces, indicates that the predecessors of Santa Paula and Sisar creeks once flowed west to join the river. Lion Creek, the beheaded portion of the original river, flows in a narrow trench incised in the broad valley floor. The principal source of water is a cienaga at the west end of the abandoned Sisar fan. Permanent streams from the Santa Ynez Mountains no longer reach the Upper Ojai Valley—they are diverted by Sisar and Santa Paula creeks.

Near the close of the Sulphur Mountain cycle the westward extension of Santa Paula Creek was captured by headward growth of the Coastal Hills part of Santa Paula Creek. The weak Pico and "Santa Margarita" shale that close around the east-plunging nose of the Sulphur Mountain anticline provide Santa Paula Creek with an erosional advantage over other streams in the Coastal Hills. These streams head in resistant Modelo shales on Sulphur Mountain. These rocks strike normal to the streams and retard the headward growth of Aliso, Wheeler, and Adams canyons.

Stages in the capture of the west-flowing drainage of the Upper Ojai Valley by Santa Paula Creek are illustrated in Figure 7. A series of terraces, the highest at about the same elevation as the Upper Ojai Valley (1500 feet) extends down the valley of Santa Paula Creek. Above

FIGURE 7.—*Stages in the capture of the Upper Ojai Valley by Santa Paula Creek*

this height Santa Paula Canyon has a gentle cross-profile, very likely a relic of the Sulphur Mountain surface.

Sisar Creek had a complex history as a result of the capture. The creek was formerly tributary to the stream crossing the Upper Ojai Valley and probably continued in its course for some time after the piracy. During this period Sisar Creek built a fan across the east end of the valley to the base of Sulphur Mountain, and was able to do so because of the absence of a large stream to remove the debris.

After Santa Paula Creek was established in its present course, an east-flowing tributary of the enlarged stream pushed westward by headward erosion along the Sisar fault. Eventually this subsequent stream reached Sisar Creek and reversed the direction of flow from west to east into Santa Paula Creek rather than into Lion Canyon and the Ventura River. As a consequence of this reversal, Sisar Creek is graded to Santa Paula Creek 500 feet below the former temporary base level of the Upper Ojai Valley. Sisar Creek, now deeply incised below the fan it once built across the Upper Ojai Valley, is actively building a steeply graded fan at a lower elevation than the one abandoned.

The Upper Ojai Valley is a vulnerable landform and cannot be expected to long survive. It is being actively attacked from three sides by streams graded to base levels 500 to 700 feet lower than the valley floor. Subsequent tributaries of Sisar Creek, extending westward along the Sisar fault system, will soon capture the eastern end of the valley. The northern side is being vigorously assailed by streams graded to the Ojai Valley. Lion Creek, graded to a base level 750 feet lower than the floor of the Upper Ojai Valley, has made the greatest progress in the destruction of the valley. The upper end of the creek, entrenched in a narrow canyon cut in Sespe shale, follows the strike of this rock for 1 mile from the west end of the valley.

Lion Creek in grading itself to the base level of San Antonio Creek has rejuvenated short tributary streams on the north side of Sulphur Mountain. Tributary streams of Lion Creek, like Big Canyon, that empty into the Upper Ojai Valley, have comparatively broad valley floors and moderately steep slopes. Streams west of the Upper Ojai Valley flow through narrower canyons with steeper gradients and join Lion Creek 250 to 300 feet lower than valleys to the east.

SULPHUR MOUNTAIN UPLAND

General statement.—The Sulphur Mountain Upland is a nearly continuous ridge, convex southward, through the center of the area, and is breached by only two large streams, Ventura River and Santa Paula Creek. The upland consists of three distinct mountains; Sulphur Moun-

tain east of the Ventura River and Red and Rincon mountains west of the river.

Sulphur Mountain.—The most conspicuous feature of Sulphur Mountain is the contrast between the late mature topography of the summit and the steep southern escarpment. The latter at times has been interpreted as a fault scarp, but is the result of rapid removal by streams of the weak "Santa Margarita" shale. The evidence for the erosional rather than tectonic origin of the escarpment may be summarized:

(1) No large-scale faulting is visible in the bedrock at the base of the escarpment. This is particularly true east of Aliso Canyon where the section, although overturned, appears complete from the Pico formation through the "Santa Margarita" into the Modelo.

(2) The height of the escarpment correlates with rock resistance.

(3) The map pattern of the escarpment is determined by subsequent streams following the strike of the "Santa Margarita" shale. Short subsequent tributaries of south-flowing Coastal Hills streams have excavated alcoves in the face of the mountain, producing a cuspate pattern.

(4) The scarp disappears west of Coche Canyon where the "Santa Margarita" strikes into the mountain, and where headward erosion by Coastal Hills streams is retarded by resistant lower Pico conglomerate.

The escarpment is highest at the cirquelike head of each of the south-flowing streams and lowest on the intervening divides. The height of the escarpment is not relatively constant,—to be expected if the cliff were a fault scarp—but varies with differences in rock resistance and with the degree of development of subsequent streams.

Streams have gentle gradients in the Coastal Hills where they cross weak Pliocene and Pleistocene rocks. Aliso Canyon drops from 1250 feet at the base of Sulphur Mountain to 500 feet at the Santa Clara Valley, 6.5 miles distant,—a gradient of 115 feet per mile. The same stream at the top of the Sulphur Mountain escarpment is at 2250 feet and drops 1000 feet to the base in 0.4 mile, a gradient of 2500 feet per mile. Because of this discordance in stream profile, streams on the summit of Sulphur Mountain have been rejuvenated, and have incised narrow canyons in their valley floors.

Red Mountain.—The most conspicuous topographic feature west of the Ventura River is the elongate dome of Red Mountain. The annular drainage pattern is only moderately well developed. Coyote Creek and Ayers Creek, its westward continuation, are the best examples of structural control of streams: both flow in strike valleys in the upper Sespe shale between the Vaqueros hogback and the central body of Sespe sandstone and shale in Red Mountain. Subsequent streams are less well-

developed south of the mountain in weak Miocene shales north of the Red Mountain thrust.

Numerous undrained depressions on the summit of Red Mountain were formed by mass movement of the Sespe sandstone down dip. These depressions are 100 to 400 feet long and 20 to 60 feet deep. Many contain permanent ponds, while others are partially filled and converted to grass-covered flats. Almost all the undrained hollows are at the east end of the mountain, and most are north of the large, now inactive slide at the head of Canada del Diablo. A large slide in Padre Juan Canyon has extended its theater-headed valley through the original crest and is now attacking the north side of the mountain where it has beheaded small north-flowing streams.

The undrained depressions are probably genetically related to landslides. They are slumped areas at the head of the slide between the area of maximum slippage and the undisturbed strata at the mountain top. The depressions are between the breakaway scarp and backward rotated blocks at the upper end of the slide.

Streams flowing directly to the ocean from Red Mountain have an advantage over those draining north to Coyote Creek at an altitude of 250 to 300 feet. Their steeper gradient enables them to erode more rapidly where they cross weak Pliocene and Pleistocene shales and sandstones at the coast. The southern streams have extended farther into Red Mountain, and the divide has migrated northward, as shown by the cirquelike heads of Padre Juan, Madranio, and Javon canyons. Two streams, Madranio and Los Sauces creeks, cut through the Red-Rincon divide, for both streams head in Rincon shale. Los Sauces Creek has extended farther north across the axis of the Ayers Creek syncline and now heads on the north side of the fold in the weak upper Sespe shale that Ayers Creek follows on the south limb.

The considerable topographic effects of the Red Mountain thrust were mentioned in the section on structure and largely result from differential erosion of resistant Miocene strata thrust over weak upper Pliocene clay shales. Overthrust Miocene rocks make a conspicuous ridge between Canada del Diablo and Canada de Rodriguez and the valleys of Padre Juan, Javon, Diablo, and Rodriguez creeks are constricted where they cross these resistant rocks.

The Modelo is a persistent ridge maker; the Rincon shale, which here overlies the Modelo in inverted sequence, is much less resistant and localizes many slides, as well as short subsequent tributaries to each of the master streams. As the Sespe formation is weaker than the Modelo, and the usually resistant Vaqueros oyster reef is very thin, or lacking, between the Sespe and Rincon formations, streams that extend northward

across the Modelo barrier north of the fault have widened their upper valleys. These streams now head in large, nearly circular theaters, constantly being enlarged by slides in the Sespe. Short subsequent tributaries in the Rincon formation are rudiments of an annular drainage pattern much less advanced than Ayers Creek north of the mountain.

Rincon Mountain.—Rincon Mountain resembles Sulphur Mountain in topographic form and geology. However, Pico sandstone and conglomerate crop out on the southeast corner of the mountain, and in the steep sea cliff east of Punta Gorda. These Pliocene strata on the north limb of the Seacliff anticline are cut off to the north by the Red Mountain thrust. Uppermost Pliocene sands and silts crop out at an altitude of 1150 feet on the main northern ridge of the mountain, and these rocks, lithologically like the "Saugus" formation at Ventura, are difficult to distinguish from the terrace gravel.

The crest of Rincon Mountain is a rounded ridge that falls off abruptly to north and south, but rises gradually eastward to the domelike summit overlooking Los Sauces Creek. This rounded crest at 2165 feet is the most conspicuous topographic feature in the coastal area, and preserves the late mature topography of the Sulphur Mountain surface.

The summit of the mountain is crossed by a few mature valleys that resemble those of Sulphur Mountain. They are being trenched by streams rejuvenated through regional uplift, especially in the weak shales above and below the Modelo on the south and north sides of the mountain respectively.

The greatest difference in the appearance of the two mountains is the marine terraces on the south side of Rincon Mountain. These rise in 9 levels to 1250-1300 feet. They have not been elevated uniformly, but are warped and increase in altitude eastward. The terrace floors are invariably buried under subaerial detritus that obscures the thin layer of fossiliferous marine sand and gravel on the wave-abraded platform.

COASTAL HILLS

The Coastal Hills include the area between Sulphur Mountain and the Santa Clara Valley and inland from the coast to the eastern boundary of the region. These low hills, 1000 to 1950 feet high, rise to nearly accordant elevations, and are crossed by streams tributary to the Ventura River, or that west of Sexton Canyon flow directly to the Pacific Ocean.

The larger streams flow normal to the strike and for the most part show little structural control. However, near Peppertree Canyon the streams curve around the eastward plunging nose of the Ventura Avenue anticline. Canada Larga nearly coincides with a synclinal axis, and Canada

del Diablo is controlled in part by a shear zone on the south limb of the Ventura Avenue anticline.

Except for the gravel benches at 1500 to 1750 feet that radiate from the high peak of upper Pico conglomerate midway between Canada Larga and Aliso Canyon (Pl. 1), the Sulphur Mountain surface is destroyed and the province is in early maturity. Elsewhere, although ridge tops are at essentially the same elevation there are no extensive flat areas on top. These accordant summits are probably a near replica of the Sulphur Mountain surface. The original elevation was reduced to its present height at a nearly uniform rate owing to homogeneity of the rocks and nearly equal spacing of the principal streams.

Gravel benches on the surviving remnant of the Sulphur Mountain surface slope outward from the 1950-foot peak and indicate that at their origin, as today, streams radiated from the Pico conglomerate ridge. Higher benches range from 1500 to 1700 feet, and a lower series at 1050 to 1250 feet is on the ridge north of Hall Canyon.

These gravel benches indicate an episode of canyon cutting followed by filling, and succeeded by the modern epicycle of deepening. Whether changes in the regimen of these streams are related to oscillations of base level, or to climatic fluctuations remains unanswered. The modern canyon epicycle is a result of late Pleistocene uplift recorded by river and marine terraces.

Numerous stream captures in the Coastal Hills are the result of lateral abstraction. The best example is the capture of the upper end of Barlow Canyon by Hall Canyon. Hall Canyon was first to tap the shale core of the Ventura Avenue anticline; the principal eastern tributary cut a deep canyon along the strike of these weak rocks. This canyon captured the headwaters of Barlow Canyon, and the stolen segment is extending eastward towards Sexton Canyon. This latter canyon is on the verge of capture, and the diversion may be complete in a few decades. Hall Canyon is 300 to 400 feet deeper than Sexton Canyon, and Hall Canyon tributaries have largely consumed the original divide. Small tributaries of Sexton Canyon have been beheaded and surviving lower portions of their valleys notch the dividing ridge.

Landslides are conspicuous features and are important agents of transportation,—the largest are shown on the geologic map, but there are many others too small to be shown. The largest are dip-slope slides and are on both flanks of the Ventura Avenue anticline (Putnam and Sharp, 1940, p. 591-600).

Earthflows are of great quantitative importance as agents of erosion. Nearly every square foot of surface on hill slopes underlain by upper Pico clay shale is in motion downslope, or has moved in the very recent

geologic past. Upper slopes of these hills are scarred by lunate breakaway scarps of earthflows, while the lower slopes are a wrinkled, hummocky surface of transported regolith. Some earthflows have deflected streams and temporarily dammed a few of the smaller canyons. Earthflows are prominent on both sides of Canada Larga and Weldon Canyon on the north limb of the Ventura Avenue anticline, and in Hall, Barlow, Sexton, and Peppertree canyons on the south limb.

VENTURA RIVER

The Ventura River flows south for 12 miles from the Santa Ynez Mountains to the Pacific Ocean 1 mile west of Ventura (Fig. 2). The maximum width of the flood plain is 0.9 mile, and only in a few places does it exceed 0.5 mile. The narrowest segment, at the east end of Red Mountain, is 500 to 600 feet wide. The average gradient of the river south of the mountains is 50 feet per mile, with the maximum slope of 70 feet per mile in the Ojai Valley.

The river has only a slight tendency to meander in the lower part of the course; elsewhere the channel is anastomosing, especially in the northern Ojai Valley. This is one of the wider parts of the flood plain which here resembles a valley train with numerous branching distributaries. In the southern Ojai Valley the flood plain is 2500 feet wide at Oakview but narrows to 700 feet at the junction with San Antonio Creek. No bedrock is exposed in this section of the channel, and the depth of fill appears to be considerable, although no measurements are available.

Bedrock crops out in the river channel at the eastern end of Red Mountain. The valley floor is only 600 feet wide and the river channel narrows to 150 feet. Rocks exposed in the river bottom are Sespe sandstone and shale and the Vaqueros oyster reef. This outcrop in the river channel, at an elevation of 200 feet and 5 miles inland, is at the place of maximum elevation of the warped river terraces.

From Sulphur Mountain southward the river is on the west side of the valley, and is actively truncating spurs, as well as rejuvenating tributary streams by shortening their length. Alluvial fans of low gradient are being constructed on the east side of the valley between Canada Larga and Canada de San Joaquin. The last 2 miles of the valley appear to be deeply alluviated, as it was probably excavated to a greater depth during the late Pleistocene low stand of sea level. However, no submarine canyon is opposite the mouth of the Ventura River (Shepard, 1941, Chart I). The reason may be the gentle gradient of the sea floor in the nearly enclosed Santa Barbara Channel. Submarine canyons appear to be restricted to the steeper continental slopes.

VENTURA RIVER TERRACES

DISTRIBUTION AND ALTITUDE

The Ventura terraces provide important evidence of the extent and nature of the deformation of the Ventura region since the close of the Sulphur Mountain cycle. Terraces on either side of the river are mostly narrow, or even isolated benches, but in the Ojai Valley they cover 7 square miles. The most complete sequence of Ojai terraces is the ridge south of Long Valley between San Antonio Creek and the Ventura River (Pl. 2, fig. 2). Terraces are almost equally well preserved, though smaller, at the west end of Sulphur Mountain and the east slope of the 1000-foot peak west of the river mouth (Pl. 3, fig. 2; Pl. 4, fig. 1).

The highest terrace south of the Santa Ynez Mountains is at 1180 feet on the west end of Sulphur Mountain, south of Fresno Canyon. The equivalent of this terrace is at an altitude of 1000 feet north of Oakview, and at approximately the same elevation west of the Ventura River mouth. The higher terraces merge with the Sulphur Mountain surface, indicating that the Ventura River was established before the late Pleistocene uplift.

Terrace elevations were determined in the field from the contour map. The contour interval is 50 feet, and the elevations are probably correct to within a half of the interval. Higher accuracy is not warranted because of the variable depth of fill.

Only one terrace may be traced essentially continuously from the Santa Ynez Mountains to the sea. This terrace, to which the name Oakview is here given, is most extensive at an altitude of 500 feet between San Antonio Creek and the Ventura River south of Oakview (Fig. 8; Pl. 3, fig. 1). It may be recognized in discontinuous patches up and downstream from Oakview. West of Fresno Canyon the elevation is 550 to 600 feet; at the western end of Sulphur Mountain, 650 feet; opposite Manuel Canyon, 500 to 550 feet; and at the river mouth the range is from 200 to 400 feet.

The Oakview terrace merges upstream with, and hence is equivalent to, the gravel-covered floor of the Santa Ana Valley. The widespread development of this surface, both as terraces along the river, and in the stripped and filled parts of the Ojai Lowland suggest that it is a strath carved during an epicycle intermediate between the present and the Sulphur Mountain cycle.

The Oakview terrace, as the correlative of the floor of the western Ojai Valley, is related in time to later drainage changes in this part of the valley. For example, the capture of the eastern half of the valley by headward growth of San Antonio Creek is contemporaneous with the

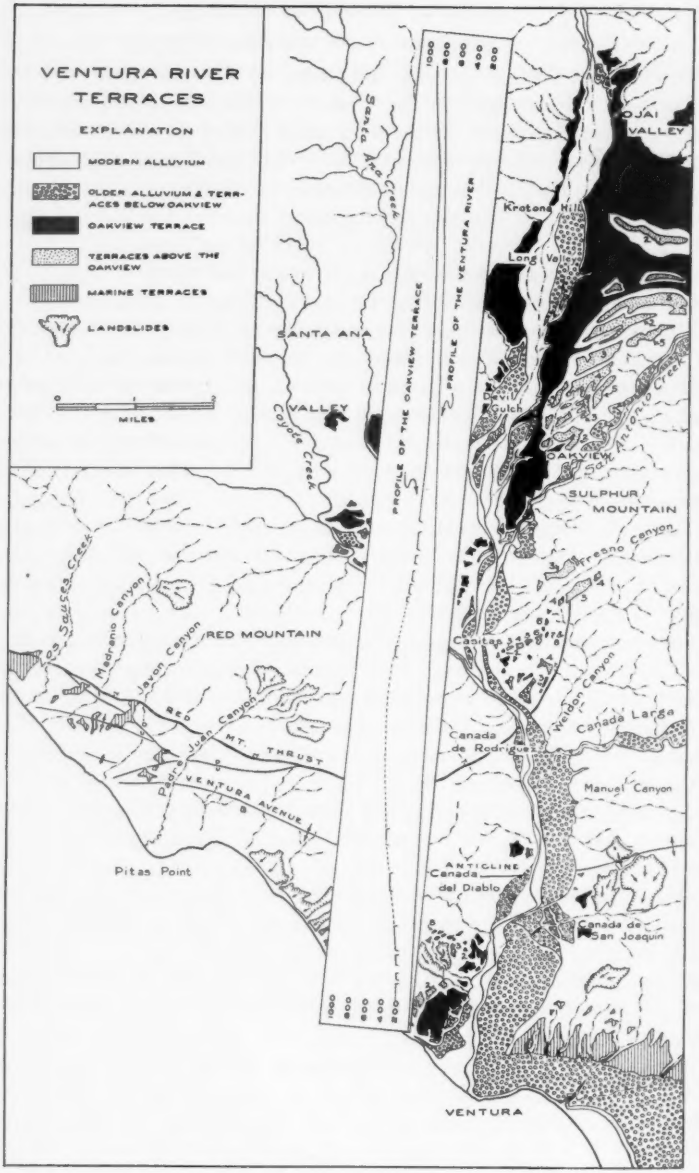

FIGURE 8.—*Ventura terraces*

cutting of the Oakview strath. The Oakview epicycle, however, is later than the capture of the Upper Ojai drainage by Santa Paula Creek, for this occurred when the higher terraces were being cut. These drainage changes are described more fully in the sections to follow.

DESCRIPTION OF THE TERRACES

General statement.—The Ventura terraces are described by districts from the Ojai Valley to the coast. As pointed out above, it is not always possible to correlate all the terraces throughout the valley. The Oakview terrace, although far from continuous, is recognized with some confidence (Fig. 8), and is taken as a datum (No. 1 terrace); terraces above it are numbered in sequence from youngest to oldest.

Ojai Lowland.—The most conspicuous terraces south of the Ojai Valley are between San Antonio Creek and the Ventura River. Seven well-defined terraces are recognized, four above the Oakview strath and two small ones below. Many small local benches occur but do not warrant separation as individual terraces.

The Oakview terrace extends as a narrow band northward from Oakview to the Ojai Valley (Pl. 2, fig. 2). The central Ojai Valley north of the Oakview mesa is divided by Krotona Hill into two parallel, east-trending valleys. They are the former courses of Señor and Horn creeks when they were tributary to the Ventura River during the Oakview epicycle before their capture by San Antonio Creek.

The altitude of the Oakview terrace increases northward from 600 feet at Devil Gulch to 700 feet at the east end of Long Valley and 750 feet at Ojai. Two sublevels are included with the Oakview terrace but not differentiated on the terrace map (Fig. 8); one above the Oakview level and the other below. The total relief separating these three terraces is 50 to 75 feet. The most conspicuous development of the sublevel below the Oakview is in the western two thirds of Long Valley. The valley evidently continued to degrade with the lowering base level of the Ventura River after the eastern end was captured by San Antonio Creek. The other sublevel mapped with the Oakview forms a moderately well-defined shoulder on the north side of Krotona Hill and is 20 to 30 feet higher than the Oakview terrace. This minor terrace is also on the south side of Long Valley and makes a bench 25 to 30 feet above the Oakview strath.

South of Oakview the terrace widens to a triangular mesa (Pl. 3, fig. 1) and rises from 500 feet at Oakview to 550 feet at the south end at San Antonio Creek. This is the northernmost of the terraces that is clearly back-tilted and it continues to rise for 2 miles farther south.

The second terrace is best preserved on the gravel-capped summit of Krotona Hill (800 to 850 feet) and on the crests of two small hills (750 to 800 feet) south of Long Valley. The wind gap at 800 feet between two hills whose altitudes are 950 and 1000 feet between San Antonio Creek and the lower 1 mile segment of Lion Creek is part of this terrace. This notch is a remnant of a valley excavated by a west-flowing stream (Pirie Channel, Fig. 9), probably tributary to the Ventura River, and very likely the predecessor of the Long Valley branch of Horn Creek before its capture by San Antonio Creek. The course of this abandoned valley was largely determined by erosion along a northeast-trending fault at the west end of Lion Mountain. The valley was approximately 1000 feet wide and followed a sweeping curve of large radius convex southward. Following its abandonment and during the Oakview epicycle the drainage of this valley was reversed and the present northward slope to San Antonio Creek was acquired. This reversal was the work of a small tributary of San Antonio Creek that extended headward through the rim of the valley.

The third terrace on the Long Valley mesa covers an area of about 0.75 square mile and slopes from 900 feet at the east end to 750 feet at the west end. The terrace is now crescent-shaped (Fig. 8), with the lowest part on the southern margin owing to erosion by a small west-flowing stream, formerly graded to the second terrace and now graded to the Oakview. This stream has lowered the southern part of the terrace 50 to 100 feet below the north rim, and isolates part of the terrace on the south bank from the main body. The isolated part is a rectangular area about 3500 feet by 1500 feet at an altitude of 750 feet.

The fourth terrace (825 to 850 feet) is near the top of the Oakview mesa about 0.5 mile east of Devil Gulch. In cross section the terrace is a flat-floored river valley with gently flaring walls. The upper and lower ends were removed by erosion, so that the valley now opens into the air at each end of the terrace. Its breadth is 3000 feet, the width of the present Ventura flood plain west of here. Terraces probably equivalent to the fourth level on the Oakview mesa are on the ridge north of Lion Canyon at 950 to 1000 feet.

The fifth and highest terrace at Oakview (950 to 1000 feet) is on the narrow ridge west of San Antonio Creek opposite Lion Canyon. The cap of gravel and boulders, some of which are 4 feet in diameter, is 20 to 40 feet thick. The conspicuous reddish knob (920 feet) on the west side of the broad valley of the fourth terrace, east of Devil Gulch, is part of the same terrace.

The interpretation of the drainage history of the Oakview area is complicated because two west-flowing streams draining the Ojai Valley

here joined the Ventura River. These tributaries were captured by headward growth of San Antonio Creek and now join the Ventura River 4.5 miles downstream at Lacrosse rather than at Long Valley. The accompanying map shows the sequence of the various stream diversions (Fig. 9).

In the first stage, at the time of the cutting of the fifth terrace, the east fork of Santa Paula Creek was tributary to the Ventura River by the Upper Ojai Valley. When this drainage was captured by present-day Santa Paula Creek at the east end of Sulphur Mountain, the volume of water in Lion Creek was sharply curtailed. At this time the Ojai Valley was connected with the Ventura River by the Pirie channel.

These same conditions of reduced flow in Lion Canyon and a western outlet for the Ojai Valley held for the cutting of the fourth, third, and second terraces. The Ojai outlet shifted northward from Pirie channel to Long Valley (Long Valley channel) and to the valley north of Krotona Hill (Krotona channel, Fig. 9) early in the Oakview epicycle. Whether or not these two outlets functioned simultaneously cannot be determined. Quite likely they did, and the northern valley was followed by Señor and Gridley creeks, while Long Valley was the outlet for Horn Creek, whose course across the Ojai Valley parallels the Santa Ana fault.

Señor and Gridley creeks may have been captured during the Oakview epicycle by a tributary of Horn Canyon along the present-day course of Stewart Canyon south of Ojai. Their present junction with Horn Creek east of Ojai was established later. The last event in the Oakview epicycle was the capture of the entire drainage of the eastern Ojai Valley by headward erosion of San Antonio Creek localized in the weak Rincon shale as far as the resistant Vaqueros sandstone where it crops out near the junction of Lion and San Antonio creeks.

Sulphur Mountain Upland.—The narrowest section of the Ventura valley is the Sulphur-Red Mountain water gap, and here the terraces are highest. All terraces here are narrow when compared to their extent in the Ojai Valley (Fig. 8). The Oakview terrace narrows from almost 3 miles to 500 feet.

Another feature of Sulphur Mountain is the terraces above 800 feet on both sides of Fresno Canyon, slightly less than half a mile east of the Ventura River. These were probably cut by the Upper Ojai stream before its capture by Santa Paula Creek, and were maintained by Lion Canyon until its diversion by San Antonio Creek.

The Oakview terrace, on the west side of the Ventura River, extends discontinuously for 1.5 miles between the junctions of San Antonio and

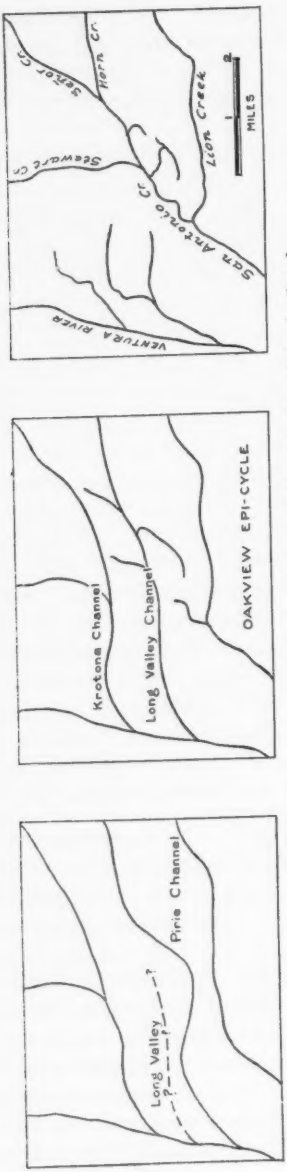

FIGURE 9.—*Stages in the capture of the Ojai Valley by San Antonio Creek*

Coyote creeks. In this distance the terrace surface rises from 500 feet at the north to 600 feet at the south. Although part of this rise is due to the Ventura River having cut farther into the terrace at the south end, nevertheless equivalent parts of the terrace are higher at the southern end; the terrace is inclined northward against the gradient of the river.

The Oakview terrace is destroyed on the west side of the river south of Coyote Creek; on the east side it survives as a gravel bench at 650 to 700 feet on Sulphur Mountain (Fig. 8). This terrace remnant is 4500 feet south of the equivalent level west of the river and is 50 to 75 feet higher. There is no certainty that the two terraces are contemporaneous, but their altitude, degree of preservation, and position in the sequence strongly indicate that the 650-foot bench on Sulphur Mountain is part of the Oakview terrace.

The Oakview terrace is also found in the valley of Coyote Creek. The two best preserved portions are the triangular area about 500 feet in altitude at the south end of lower Santa Ana Valley and the crest of the narrow ridge (550-600 feet) east of the junction of Ayers and Coyote creeks. This terrace, like the Oakview terrace on the Ventura River, slopes north against Santa Ana Creek.

Terraces above the Oakview level are scattered, gravel-covered benches, chiefly confined to the west end of Sulphur Mountain south of Fresno Canyon. The vertical interval separating them is slight, and it is difficult to correlate individual terraces. For this reason, no attempt is made to correlate the Sulphur Mountain terraces, other than the Oakview terrace, with those of the Oakview mesa.

Seven benches are recognized above the Oakview terrace on the Sulphur Mountain ridge separating Fresno and Weldon canyons. These are at the following elevations: (1) 650-670 (Oakview), (2) 800, (3) 880-900, (4) 950, (5) 975-1000, (6) 1100, (7) 1130, (8) 1150-1180 feet. The seventh and eighth terraces are on the dome-shaped summit of the ridge. Gravels of the other terraces cap the three prominent ridges between the summit and the Ventura River. Of these terraces the fourth and fifth are broadest and cover most of the flat-topped portion of the northern and southern of the three ridges. The fourth and fifth terraces are better preserved than the lower benches since they are cut in resistant limestone and shale of the Modelo formation, while the lower terraces are in the weak Rincon shale.

Two conspicuous terraces below the Oakview at Sulphur Mountain are at 400 to 450 and 475 feet. The 400-foot terrace is more extensive and makes a broad bench northwest of Weldon Canyon. Other minor terraces are at varying elevations above the flood plain.

Fresno Canyon terraces.—The origin of the boulder-covered benches on either side of Fresno Canyon is one of the physiographic problems of the Ventura region. These terraces are 1.2 miles east of the junction of the canyon with the Ventura River and are inclined north like the river terraces. The average altitude of the terrace on the right bank is 800 feet in the trough of the former valley, while the left bank terrace slopes northward at an average gradient of 4 per cent from 950 feet to 850 feet. In cross section these terraces appear as a broadly flaring valley: the eastern side rises gradually to the Sulphur Mountain summit, the western to the crest of the terraced hill at the west end of the mountain (Pl. 3, fig. 3).

The Fresno Canyon wind gap seems to be equivalent to the second terrace above the Oakview at the west end of Sulphur Mountain. That the Fresno Canyon terraces were not cut by the Ventura River, unless it divided around an island, is indicated by terraces that extend up the eastern slope of the peak between Weldon and Fresno canyons. There are at least four distinct terraces at elevations equivalent to the fourth, fifth, sixth, and seventh, levels on the Ventura side of the ridge.

Quite likely the abandoned valley at Fresno Canyon was occupied during the Sulphur Mountain cycle by the Upper Ojai River before its capture by Santa Paula Creek. The Upper Ojai River probably joined the Ventura River near the mouth of Weldon Canyon. All terraces south of Fresno Canyon were removed by erosion, as this area is south of the Red Mountain thrust and the rocks exposed are weak shales of the upper Pico. Present-day San Antonio Creek has destroyed all trace of the former drainage system between Fresno Canyon and the Oakview Mesa. This occurred after the capture of the lower part of Lion Canyon by San Antonio Creek (Fig. 10).

Coastal Hills.—River terraces are not prominent in the Coastal Hills. The most conspicuous,—probably the correlative of the Oakview—is west of the river and opposite Manuel Canyon. The Oakview level is also prominent on the only extensively terraced hill in the province, the 1000-foot peak west of the river mouth. Other fragmentary benches and scattered gravel patches occur, but cannot be correlated.

The correlation of the Oakview terrace at the type locality with scattered remnants in the Coastal Hills is made partly on the basis of comparable development and preservation, and partly because of the profile (Fig. 8) that can be drawn through them. From the southernmost remnant of unequivocal Oakview terrace west of the river at Fresno Canyon, it is nearly 4500 feet to the remnant on Sulphur Mountain at 670 feet altitude. A gap of 2.3 miles intervenes to the prominent mesa at 500-550 feet alti-

tude across the river from Manuel Canyon (Fig. 8). A higher terrace at 600 feet and a lower at 350 feet also occur here.

Terraces east of the river at Canada de San Joaquin are cut in the same resistant Pico sandstone on the south limb of the Ventura Avenue

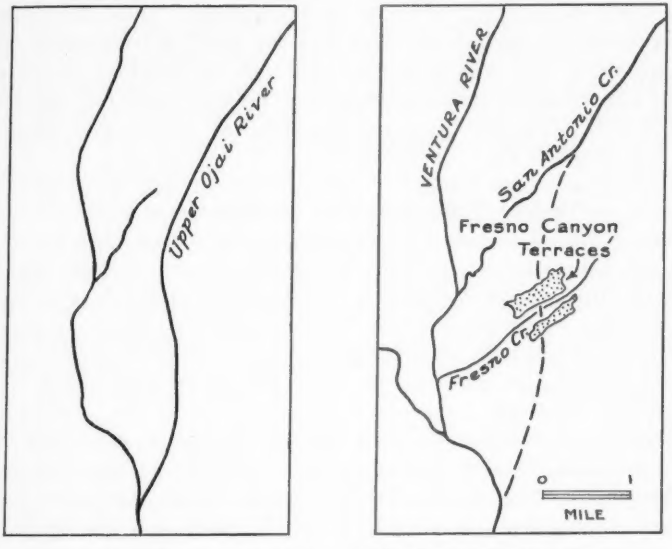

FIGURE 10.—*Stages in the history of the Fresno Canyon terraces*

anticline that preserves the terraces opposite Manuel Canyon on the north limb. The Oakview terrace makes a bench at 500 feet altitude north of Canada de San Joaquin. There are three terraces below the Oakview, at elevations of 300, 200, and 125-150 feet.

No higher terraces are recognized certainly in the Ventura Avenue oil field. However, on the ridge north of Canada de San Joaquin some gravel is found at 600 feet and at 750 feet. Near the crest of the ridge at 900 to 1500 feet is a nearly level area 2000 feet long by 500 feet wide that slopes gently toward the Ventura River. Although no gravel is present, this flat-topped ridge appears to be part of the broad Ventura valley cut during the Sulphur Mountain cycle. It is probably the correlative of the 1180-foot terrace on Sulphur Mountain, as well as the high-level gravels surrounding the 1950-foot peak east of Canada Larga.

Terraces west of the Ventura River and south of Canada del Diablo (Pl. 3, fig. 2; pl. 4, fig. 1) rise in nine levels from the Ventura River to

1030 feet. The three lower levels are the most conspicuous; higher levels are less readily recognized and are generally gravel-covered spurs and ridge tops. Unquestionably they are terrace remnants, for the gravel, which rests on clay shale, contains boulders several feet in diameter. Higher terraces are less certainly recognized in the southern third of the area where they are cut in the "Saugus" formation. Here the lithology of terrace gravel and bedrock is identical, and gravel of the upper terraces is spread over the lower, so that terraces from the fourth to the seventh make a gravel cap that slopes continuously from an altitude of 900 down to 600 feet. This gravel cap is on the nearer of the two dark ridges at the left margin of Figure 2 of Plate 3.

Terraces are most completely preserved on the first ridge south of Canada del Diablo. The terrace below the Oakview is at an altitude of 150 feet. Others, including the Oakview, are at the following altitudes: (1) 300-400 (Oakview), (2) 510, (3) 600, (4) 700, (5) 750, (6) 825, (7) 900, (8) 1010 feet. Of this series the fourth and fifth terraces are separated by so slight an interval that they may be regarded as the same, so there are only eight clearly differentiated terraces. Gravel benches are also found between this ridge and the coast, and on the main divide to an altitude of 1030 feet.

Terraces south of Canada del Diablo are inclined seaward at a slope greater than the present gradient of the Ventura River. This inclination is best demonstrated by the Oakview terrace, the most completely preserved of the series. The northern end of the terrace is at 300 to 450 feet, while the southern is 150 to 200 feet at the edge of the modern sea cliff. The average slope of the terrace is 100 feet per mile, compared to a gradient of 40 feet per mile for the seaward portion of the Ventura flood plain.

Summary.—The purpose of the preceding description of the Ventura River terraces is to demonstrate that not only is the Ventura valley terraced, but also that the terraces are arched and reach their maximum elevation at the Red Mountain fault. To the extent that the Ventura River has maintained its course against this minor uparching it is antecedent, which of course does not imply that the entire history of the river is to be thus explained.

The origin of the Ventura River extends back to the Sulphur Mountain cycle when the river probably established its present course by headward growth and capture of an earlier drainage system, possibly in the same way as Santa Paula Creek diverted the west-flowing drainage of the Upper Ojai Valley. The adjustment to structure of the river in the Sulphur Mountain area supports the belief that it extended itself across the Sulphur Mountain Upland by headward erosion largely in the Rincon

shale rather than by superposition. The original captured drainage cannot be reconstructed, as the piracy occurred during the Sulphur Mountain cycle and all evidence of terraces and wind gaps is destroyed. Probably the Ojai Lowland then drained westward by a route paralleling East and West Casitas passes.

Deepening of the Ventura River valley kept pace with the pulsatory uplift recorded by the terraces. Finally, the terraces were arched and this movement is presumably still continuing. The post-Oakview terraces are warped, the most conspicuous of these is 150 feet above the river at Sulphur Mountain. The narrowness of the valley and flood plain, and the bedrock outcrops in the Sulphur Mountain section doubtless result from warping, at least in part.

COASTAL PHYSIOGRAPHY

MARINE TERRACES

General statement.—Aside from the low wave-cut platform at the base of the modern sea cliff, marine terraces are restricted to two sections of the coast: the first from Pitas Point 7 miles west to the Carpinteria Plain, and the second from the Ventura River east to the Santa Clara River. Terraces are prominent west of Pitas Point and are best developed on Rincon Mountain where they reach an altitude of 1300 feet. Elsewhere, the number and altitude are difficult to determine, for they occur as discontinuous gravel patches on ridge tops. Terraces are obscure behind Ventura because their gravels are nearly identical with the underlying "Saugus" formation.

Marine terraces near Rincon Mountain (and perhaps elsewhere) have been tilted so that they slope westward towards the Carpinteria Plain. Almost all are cut by faults which offset the terrace floor—in one case as much as 30 feet. Many smaller faults cutting the terraces may be seen in road cuts, particularly on the ridge between Javon and Madranio canyons and on the reservoir road behind Ventura.

Recognition of individual terraces is equally difficult in the Ventura region as in the San Pedro Hills (Woodring, 1935, p. 292-305). The greatest difficulty is caused by alluvial fans constructed on the terrace surface,—fan gravels 150 feet thick are revealed in several canyons with the base still unexposed. A single alluvial fan may bury several terraces so that an originally benched surface is converted into a continuous slope. The surface of the alluvial gravels is convex at the principal canyon mouths; the entire terrace may acquire an undulating profile like that of a piedmont alluvial plain. It is rarely possible to estimate the depth of fill, and the empirical method used by Davis (1933, p. 1057) yielded no satisfactory determination of the thickness of subaerial detritus.

These difficulties of terrace recognition and correlation frequently render indeterminate both the number of terraces present, and the altitudes of individual terraces. Altitudes of their alluviated surfaces were determined in the field from the Ventura quadrangle map and are probably accurate to one contour interval.

The marine part of the gravel blanket on the terrace floor is quite thin,—rarely 4 and usually less than 2 feet thick. The marine gravel of cobbles and boulders is set in a matrix of cross-bedded, and generally iron-stained sand. Most of the boulders are Eocene sandstone, or concretionary Miocene sandstone, but quartzite, plutonic, and volcanic igneous rocks, and cherts also occur. Limestone and siliceous shale from the Modelo formation are abundant in the gravels of the western part of the coastal section.

Rincon Mountain.—A series of conspicuous benches on the seaward slope of Rincon Mountain reaches an altitude of 1250 to 1300 feet. The fourth terrace, at an altitude of 550 feet, is prominent at the eastern end of the mountain and is the broad platform at the top of the sea cliff in Figure 1 of Plate 5. The westward inclination of this terrace is apparent in the photograph,—discounting the fact that the sea has cut farther back in the terrace at the eastern end than behind Punta Gorda, at the left margin of the figure. The highest marine terrace is the comparatively broad bench at the base of the steep cliff below the summit ridge, and is directly above the center of the photograph.

Marine terraces on Rincon Mountain are most readily recognized, and have the least detrital cover on the long ridge that rises eastward from the mouth of Rincon Creek to the mountain crest. Nine terraces are on this ridge at the following elevations: (1) 70, (2) 200, (3) 300, (4) 425-450, (5) 500-525, (6) 600-650, (7) 750, (8) 950, (9) 1075-1150. The sixth is the most extensive, while at the east end of the mountain the fourth is the best preserved; all terraces below it have been destroyed in cutting the present coast. At the west end of the mountain, terraces above the sixth cannot conclusively be proved marine, as no fossils or mollusk-bored sandstone boulders were found. These terraces, covered with rounded cobbles and boulders of Eocene and Sespe sandstone, were probably cut by Rincon Creek. However, they were no great distance inland when cut, and thus are a reasonably close measure of the position of sea level at the time of their origin.

The highest (ninth) terrace may be correlated with fair certainty the length of Rincon Mountain. It is shown on the Ventura quadrangle as the level area inclosed by the 1250-foot contour south of the principal canyon, and also by the bench between 1150 and 1300 feet just north of

the canyon. A pholad-bored sandstone boulder at 1200 feet on this terrace shows it to be marine. If this terrace is the equivalent of the ninth terrace at 1075-1100 feet on the northern ridge, the increase in elevation southeastward to 1250-1300 feet indicates that these terraces are tilted northwestward. This northwestward inclination is confirmed by the fourth terrace, which at the southeast end of Rincon Mountain is at 550 feet, while at the northern end it is 425-450 feet.

The westward component of this slope is more apparent farther west. The second, or 200-foot terrace may be recognized at this elevation on both sides of Rincon Creek, but slopes westward for 3 miles, and disappears below sea level at Carpinteria estero. The Carpinteria tar pits are on this terrace (Pl. 5, fig. 2).

Faulting, as well as warping, complicates the recognition and correlation of individual terraces. Three large faults cross the Carpinteria terrace and strike into Rincon Mountain. Well logs show that these faults dip south and thrust Miocene over Pleistocene strata. The northern fault has the greatest displacement and most conspicuous scarp: 100 feet high and approximately 1.5 miles long. There is a pronounced sag pond north of the cliff. The scarp of the central fault is 50 to 75 feet high and that of the southern 30 feet. The southern fault is exposed in a road cut on the coast highway west of Rincon Creek and reveals Miocene shale thrust over Pleistocene dune sand (Kew, 1924, Pl. 13, fig. B).

The three faults at Carpinteria strike across the terraced western ridge of Rincon Mountain and are responsible for undrained basins and low, east-facing scarps that interrupt the fourth, fifth, and sixth terraces. A probable fault scarp 30 to 50 feet high crosses the fourth terrace at the east end and may be related to the Red Mountain thrust. The scarp faces south, and trends diagonally across the terrace from the landward margin at Los Sauces Creek to the edge of the modern sea cliff half a mile north of Punta Gorda. The low scarp makes acute angles both with the present coast line and with the coast line at the time the terrace was cut. This rules out a wave-cut origin, and as the cliff follows the trace of the Red Mountain fault, the likelihood is strong that this is a tectonic rather than erosional feature.

Los Sauces to Pitas Point.—Marine terraces are on each of the three ridges separating the four canyons of Los Sauces, Madranio, and Padre Juan creeks. These ridge tops (average elevation 700 to 800 feet) are now so narrow that the original flat terrace surface is destroyed. Only the remnant of nearly horizontal gravel resting on strongly folded Pico sandstone and shale records the former existence of terraces. The basal 4 feet or less is marine and contains pholad-bored or barnacle-encrusted

sandstone boulders in cross-bedded sand. Many of the sand lenses, particularly on the ridge between Javon and Madranio canyons, contain fossil shells. These are listed in the section on stratigraphy.

The nonmarine part of the gravel is occasionally as thick as on Rincon Mountain, and individual terraces are less readily distinguished because of the destruction of the constructional surface by erosion. For these reasons, and because the remnants are widely separated, higher terraces cannot confidently be correlated either among themselves or with the Rincon series. Except for a small bench at 150 to 200 feet on the north side of Javon Canyon, all levels below the fourth are destroyed, and the fourth terrace survives as a narrow shelf at 500 to 550 feet on the Los Sauces-Madranio and Madranio-Javon ridges. The main gravel capping on each of the three principal ridges may possibly be equivalent to the fifth and sixth terraces at the north end of Rincon Mountain.

Faulting of these terraces is noteworthy. The Red Mountain thrust truncates each terrace at the landward margin. The terraces are also cut by the south-dipping reverse fault south of the Red Mountain thrust (Fig. 8). Renewed movement has apparently occurred on each of these faults, as well as a score of minor faults, many of which are exposed in road cuts. On the Madranio-Javon ridge the terrace floor is stepped down 20 to 30 feet by three small normal faults, and on the downthrown side of the westernmost fault is tilted seaward 20° to 25°.

A small bench with pholad-bored sandstone boulders, probably a remnant of the fourth terrace, is at 500 feet altitude on the seaward end of the Madranio-Los Sauces ridge. This terrace is at 500 to 550 feet on the Madranio-Javon ridge; the same altitude as the open-ended valley that crosses the ridge top. This is the abandoned valley of a small stream parallel to Madranio Canyon and graded to sea level at the time the fourth terrace was cut. With renewal of uplift Madranio Creek deepened its canyon more rapidly than the smaller stream and captured it.

There is a southward increase in elevation of the gravel surface on the three main ridges. The average altitude is 700 feet on the Los Sauces-Madranio ridge, 750-800 feet on the Madranio-Javon ridge, and 850 feet on the Javon-Padre Juan divide. Several terraces may be present on these ridge tops, but the low sea cliffs separating them are effectively hidden by the gravel. If the gravel caps are approximately equivalent, this increase in elevation supports the belief that the terraces are tilted northwest.

The western peak of the terraced portion of the Javon-Padre Juan ridge is 923 feet high, the eastern 850 feet. The western end stands about 70 to 75 feet above the eastern, either because it was a low island at

the time the eastern end was being cut, or, more likely, because of deformation. Outcrops are lacking to test these alternatives. A fault passes through the saddle north of the high point which is just north of the axis of the Ventura Avenue anticline. Terrace gravels on the 923-foot peak appear to be tilted north. This tilting of the terrace, and its superior elevation, may result either from renewed folding of the anticline or Recent displacement on the fault.

Pitas Point to Ventura River.—Doubtless because of rapid retreat of the shore line in the weak Pliocene and Pleistocene rocks here exposed, only two minor marine terraces are preserved between Padre Juan Canyon and the Ventura River. The first is at an altitude of 925 to 950 feet on the 1001-foot hill 1 mile east of Padre Juan Canyon and half a mile inland from the coast.

The second terrace, at an elevation of 60 to 70 feet, probably equivalent to the 70-foot bench at Rincon Creek, is a boulder and cobble covered bench at the base of the sea cliff 4 miles northwest of the Ventura River.

Ventura River to Santa Clara Valley.—Marine terraces are cut in the seaward-dipping "Saugus" formation behind Ventura. These terraces are difficult to recognize because of the similar lithology of bedrock and terrace gravel. The "Saugus" sand and gravel erode readily, so most of the higher terraces have been destroyed, and the lower are concealed under detritus from the higher slopes.

Terrace gravel is best distinguished from the "Saugus" by the attitude; dips in the "Saugus" range from vertical to 40°, in the terrace gravel from nearly horizontal to 20°. The sandy matrix of gravels below the 500-foot terrace is usually lighter buff than the "Saugus" sand, and the weathered zone may be pinkish red. The terrace form, best preserved on the lower terraces, is practically destroyed above the 500-foot terrace, and boulder-covered shoulders on the principal seaward-facing ridges are the only evidence of their former presence. Not much weight can be given this because of the possibility that the boulders may be from the "Saugus" conglomerate.

The gravel is 60 to 70 feet thick at the seaward end on the terrace behind the Ventura reservoir; up to 140 feet at the landward end; only the basal 5 to 10 feet is marine. Boulders on this basal gravel are from the underlying vertical "Saugus" conglomerate, and were deposited by waves on the landward side of the conglomerate outcrop.

The best-preserved marine terraces are on the ridge immediately east of the Ventura River flood plain. The first terrace, 2100 feet wide, on which the business district of Ventura is built is nearly level and at an

average elevation of 15 to 20 feet. Higher terraces are at altitudes of (2) 75-80, (3) 200, (4) 350-400, (5) 500-550, (6) 650, (7) 750-777. Two doubtful terraces at about (8) 800 and (9) 900 feet are on the ridge west of Hall Canyon.

The 80-foot terrace is a conspicuous bench at the east margin of the Ventura River valley. Both this terrace and the lower one are largely buried to the east by the alluvial fan at the mouth of the first canyon west of Hall Canyon. This stream and the one in Hall Canyon have incised arroyos 50 feet deep across the fan,—probably a result of the shortening of stream length through wave attack. Before the present cycle of a prograding shore line was started by the northward shift of the Santa Clara River mouth, the shore line was retrograding at Ventura. The sea cut a 35- to 50-foot cliff in the alluvial fan during the retrograding phase.

The striking similarity in lithology of terrace gravel and bedrock is well shown where the floor of the 200-foot terrace is conspicuously displayed near the top of a large excavation behind the Ventura County Courthouse. Both are unconsolidated sand, gravel, and cobbles—the "Saugus" inclined at nearly 50° and the terrace gravel almost horizontal. Gravel on this same terrace dips seaward 18° to 20° where it is exposed in cuts on the road west of the courthouse that leads to the Padre Serra cross.

The 350- to 400-foot terrace is the flat ridge top on which stands the Serra cross. The bedrock floor between the 200- and 350-foot terraces is revealed in cuts along the road to the city reservoir. These two levels, instead of being separated by a single sea cliff, are separated by four minor steps. Each of the risers of these steps is 20 to 30 feet high and is only a more steeply inclined section of the bedrock floor. The close spacing of these minor benches, the small vertical interval separating them, and the unconsolidated bedrock in which they are cut confirm the belief that the late Pleistocene emergence was probably accomplished by short and rapid uplifts, rather than by ones of large magnitude separated by long time intervals.

The terraces are faulted, and several vertical and south-dipping, high-angle reverse faults that offset both bedrock and gravel may be seen on the reservoir road. The maximum throw is about 8 feet, and the average is 2 to 3 feet. Displacement on these faults is probably the result of Recent movement on larger Pleistocene faults, as "Saugus" dips change greatly on one of them. At the junction of the road to the cross with the reservoir road, two small normal faults, one of 2-foot, the other of 5-foot throw, tilt the bedrock floor 10° to 15° seaward.

Above 500 feet the terrace landform is largely destroyed; below this level enough of the terrace survives to be recognized at a distance. The same situation exists on the west side of the river, as may be seen on Figure 1 of Plate 2. Perhaps the rate of uplift was not uniform, and the 500-foot terrace was cut during a still-stand that lasted long enough for the upper terraces to be nearly destroyed by erosion. The pronounced two-story profile of the principal streams in the Coastal Hills supports this belief.

Marine terraces appear to extend east to Barlow Canyon, although the actual eastern limit is indeterminate. South-trending ridges are terraced along the margin of the Santa Clara Valley, and these terraces appear to result from downcutting by the river. Like the terraces behind Ventura, they are cut in the "Saugus," and are equally difficult to identify. They may be recognized to 800 feet altitude, and the 500-foot terrace behind Santa Paula and between Wheeler and Aliso canyons is especially prominent.

VENTURA SHORE LINE

General considerations.—The development of the Ventura shore line is controlled by the direction of the prevailing wind and current and by the unequal resistance of the rocks exposed to wave attack. The asymmetric form of the three major headlands of the northwest-trending coast is shown on Figure 2. Their asymmetric shape, with a long northern tangent, and short northeast-trending re-entrant is the indirect result of the prevailing westerly wind.

This sea breeze is strongest in summer when the Santa Clara and Ventura valleys connect heated inland areas with the coast. The littoral current flows eastward as a result, and the usual direction of travel for wave crests is N. 50° E. Waves running in this direction parallel the longer stretches of beach and approach the shorter tangents east of each point obliquely. The angle between shore and wave front is usually 20° to 30°, and the shore line in each section of oblique approach is actively eroding.

Between Rincon Point and the Ventura River the beach is a thin veneer of sand, and bedrock crops out for considerable distances. The widest exposure is east of Pitas Point, and for nearly a mile the steeply dipping Pico sandstone and shale is truncated to a smooth platform. Other areas of exposed bedrock are: (1) at the base of the Carpinteria sea cliff, (2) east of Punta Gorda, (3) midway between Punta Gorda and Pitas Point, and (4) the beach for 1 mile west of the Ventura River. This last is unusual, for steeply inclined "Saugus" crops out in this section. The beach is covered with boulders supplied by the erosion of the "Saugus" conglomerate, rather than by the Ventura River.

The widest beach is at Pierpont Bay (Pl. 2, fig. 1, foreground), southeast of the Ventura River. This bay is the shallow re-entrant between the deltas of the Ventura and Santa Clara rivers, and is supplied by sand chiefly from the Ventura River. The beach is prograding after having retrograded during most of the Recent. The shore line is now as much as 2500 feet seaward of the cliff, and the intervening beach was built because of the northward migration of the Santa Clara River across its flood plain and delta. This northward shift is responsible for the strongly convex shore line at the river mouth. This convexity forces the littoral current to deposit most of the Ventura River sand in the bight between the two deltas. Comparison of U. S. Coast and Geodetic Survey field sheets for 1855 and 1933 shows about 580 feet of seaward advance of the shore line in 78 years (Fig. 11).

The northward migration of the Santa Clara River is shown by its present truncation of fans graded to it from the north side of the valley. Streams crossing these fans are incised in arroyos 50 feet deep. Further evidence of the northward displacement is afforded by the Hueneme submarine canyon, which is 315-320 fathoms deep (U. S. Coast and Geod. Survey Chart 5101-A) at the seaward end and about 7 miles southeast of the river mouth at the landward end (Fig. 11). The northward shift of the river mouth occurred in the Recent, if the canyon was excavated during the Pleistocene low stand of sea level.

Seacliff bench.—The last episode in the history of the Ventura coast was a slight emergence, shown convincingly by the wave-cut platform exposed on the coast highway north of Pitas Point. The marine origin of this platform is demonstrated by a layer 2 to 5 feet thick of gravel, cobbles, and boulders embedded in shell-bearing sand. Many of the boulders are riddled with pholad holes and some are barnacle encrusted.

The platform is exposed in a highway cut for 300 feet at an average altitude of 15 to 16 feet above sea level. The altitude of the exposed bedrock is 21 feet, 1.5 miles southeast of Seacliff station in culvert 390.06 of the Southern Pacific Railroad. This last exposure is east of the highway outcrop, and the inclination of the terrace surface between the two is 9°. If the terrace surface is projected landward beneath the cover of subaerial detritus, its altitude at the base of the former sea cliff is probably 45 feet. The surviving remnant of the platform is 300 feet wide, and the cliff at the landward margin is 115 to 120 feet high.

The name Seacliff bench is proposed for this emerged wave-cut platform because of its excellent preservation near Seacliff station on the Southern Pacific Railroad. Here the bench is 1200 feet wide; the mini-

mum width is 300 feet where the bench is on the up-current or northern side of each of the principal points.

Additional evidence that the Seacliff bench is an emerged wave-cut

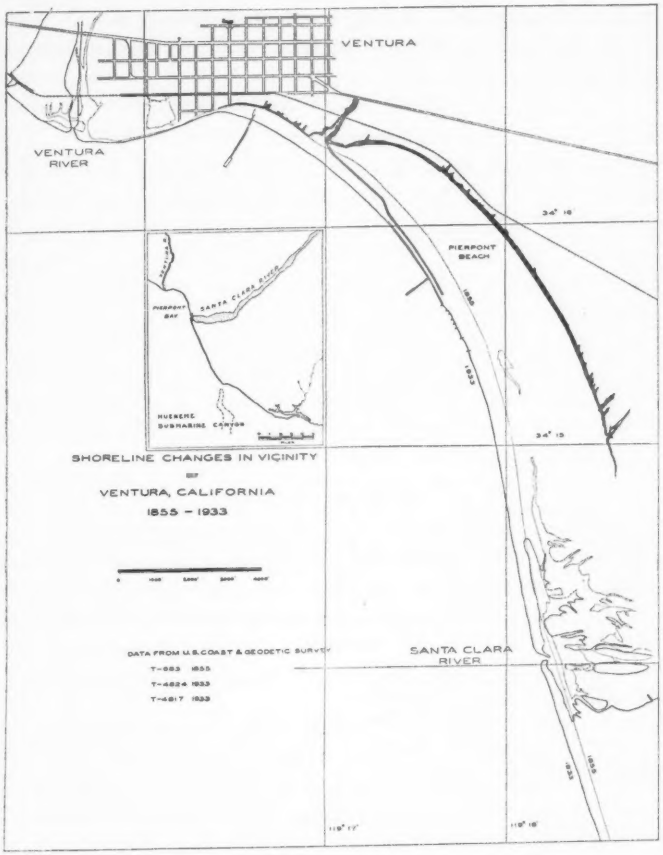

FIGURE 11.—*Shoreline changes at Pierpont Bay, 1855-1933*

platform is provided by pholad borings in a ledge of Pico sandstone that crops out landward of Mussel Rock at Punta Gorda. The holes are 5 to 7 feet above the reach of the highest tides. Though slightly enlarged by weathering, these pholad holes are identical with ones now being bored in the rock under water.

It is not possible to tell from this limited occurrence whether the Seacliff bench resulted from lowering of sea level or rise of land. Near-by marine terraces rising to 1300 feet indicate considerable vertical uplift in the late Pleistocene. It may be that the wave-cut platform is the product of the two processes and is a slightly elevated eustatic bench. This accounts for a greater elevation of the platform here than is recorded in other regions.

SUMMARY AND CONCLUSIONS

GENERAL STATEMENT

The Ventura landscape is the product of stream and wave erosion of folded and faulted shales, sandstones, and conglomerates of varying resistance, that following their deformation in the medial Pleistocene, were reduced to a surface of late maturity. Later uplift, indicated by the river and marine terraces, began the present cycle of erosion which has now reached early maturity. The evolution of the present landscape is largely controlled by structure and by differential resistance of the rocks. In the following section the stages in this evolution are discussed in chronological order: the pre-Pleistocene history is briefly considered in order to provide a background for the Quaternary history; the Pleistocene is subdivided into four major episodes.

PRE-PLEISTOCENE HISTORY

The pre-Tertiary history of the region is not known, but from late Eocene through early Pleistocene the record is primarily depositional. All rocks are marine, except the Sespe and upper "Saugus" formations, and are essentially conformable with gradational contacts. In general, except for the Oligocene (?) emergence, the Ventura embayment has narrowed and shoaled from its probable maximum extent in the Eocene to its minimum in the Pleistocene.

In the medial Tertiary the Ventura Basin was flooded following the temporary withdrawal of the sea during the deposition of the Sespe formation. The Vaqueros, Rincon, and Modelo formations afford an excellent example of progressive overlap. The "Santa Margarita," Pico, and "Saugus" formations were deposited in a long, narrow embayment whose axis had about the same trend as the present Santa Clara Valley. The northern part of the region was deformed during the early Pliocene, as is indicated by the abrupt base of the Pico conglomerate in the central part of the area. The large boulders and torrential bedding indicate that the conglomerate is probably premonitory. The lenticular conglomerate grades southward into sandstone and eventually, if the entire section could be revealed, probably into shale.

The environment in which these sediments accumulated was presumably not unlike the present Santa Barbara Channel. The conglomerate was deposited under conditions like those at the mouths of the Ventura River and Rincon Creek, where fan-shaped boulder deltas grade seaward and along shore into sand and finally into mud. As in the modern analogs, more sand was deposited on the downcurrent side of these Pliocene boulder deltas—most of the Pico conglomerate lenses grade more rapidly into clay shale westward than eastward.

Present information shows that during the Pliocene the central part of the Coastal Hills was the site of most active deposition. This part of the basin was opposite the mouths of one or more large streams that carried boulders and cobbles from all the pre-Pliocene formations. Neither the sedimentary structure nor grade size indicates that the Pico formation was deposited in deep water. The writer believes that sediment brought from a region of moderate relief was deposited in a subsiding embayment of about the dimensions of the Santa Barbara Channel. That the hinterland of the Ventura Basin had less relief in late than early Pliocene is shown by the fine grain of the upper Pico, which is dominantly clay shale except in the central Coastal Hills.

PLEISTOCENE HISTORY

General considerations.—The Pleistocene in the Ventura region is remarkable for the number and magnitude of the events occurring. There were four major episodes: (1) deposition of 4000 to 6000 feet of marine and nonmarine sediments; (2) deformation of the rocks of the region; (3) erosion, to late maturity; and (4) vertical uplift to an average of 1000 feet, and development of the present landscape by renewed stream and marine erosion.

Deposition.—During the deposition of 4000 to 6000 feet of "Saugus" formation the Ventura basin was more constricted than in the Pliocene. The "Saugus" sedimentary cycle began with the deposition of boulders, cobbles, and coarse sand, in contrast to fine-grained sediment deposited at the close of the preceding cycle.

Inferred conditions of deposition of the "Saugus" formation are duplicated today by the seaward portion of the Santa Clara flood plain and the Oxnard Plain. This area south of Ventura appears to be subsiding, but the volume of sediment supplied by the Santa Clara River is great enough to keep the trough filled and the shore line at the river mouth is prograding. If a longitudinal section were cut through the flood plain and delta it would show a wedge of continental sediments thickening landward and overlying marine sediments that conversely thicken seaward.

There was a southward shift of the axis of the sedimentary basin in the Pliocene and Pleistocene. During the Pliocene the site of maximum accumulation was the present central part of the Coastal Hills. The reason the axis shifted southward in the Pleistocene was the beginning of deformation in the northern part of the region. The Santa Clara Valley, narrowed and reduced by the medial Pleistocene deformation, is the surviving landward remnant of the Ventura basin and the Santa Barbara Channel is the remaining submarine portion.

Deformation.—Deposition of the "Saugus" formation was ended by the medial Pleistocene orogeny that developed most of the structures of the region. Folds produced during the deformation trend nearly east-west and either plunge east or are doubly plunging. Axial planes are mostly inclined, and in different parts of the same fold may dip in opposite directions, as in the Ventura Avenue anticline.

The larger faults are high-angle reverse faults, and many, like the San Cayetano and Red Mountain thrusts, dip north. Kerr and Schenck (1928, p. 1101) cited this evidence to show that stresses responsible for deforming the region operated from inland towards the coast. However, there are several south-dipping reverse faults, such as the Sisar fault and some of these are still active, particularly near Carpinteria. Over most of the region the last overthrusting was from the north, as illustrated at the east end of Sulphur Mountain and the vicinity of Padre Juan Canyon where earlier south-dipping reverse faults are overridden by the San Cayetano and Red Mountain thrusts, respectively.

The post-Pliocene age of the principal deformation of the southern Coast Ranges has been recognized for some time (Lawson, 1893, p. 115-160), and in recent years Stille (1936, p. 849-880), Reed (1933, p. 285-286, 148-150), and Eaton (1928, p. 111-141) have emphasized the medial Pleistocene age and importance of this orogeny.

The post-"Saugus" orogeny is only one of several diastrophic periods in the Pliocene and Pleistocene of southern California. Evidence for other periods of deformation in the Ventura region is: an angular unconformity between Miocene and lower Pleistocene rocks at Rincon Creek; the pronounced lithologic break between the "Santa Margarita" and Pico formations and between the upper Pico and "Saugus." The deposition of the "Saugus" formation may be contemporaneous with the beginning of the principal orogeny north and east of the western Ventura basin.

It may be concluded that throughout the Tertiary and Quaternary, deformation has been concurrent with sedimentation in different parts of the basin at different times; but the magnitude of the medial Pleistocene orogeny has concentrated the most attention on it. Diastrophism is still

active, as indicated by warped and faulted terraces, tilted gravels, and Recent fault scarps.

Erosion.—The sea was largely driven from the Ventura basin during the post-"Saugus" orogeny, and the former embayment became a land area of moderate relief. The actual magnitude of this relief, and the drainage pattern cannot now be determined. The relief probably was unimpressive, as erosion proceeded contemporaneously with deformation. At the close of the period of most active diastrophism the streams may have been consequent on the structures, thus flowing east or west, rather than south as do the principal modern streams.

The post-deformation cycle of erosion progressed to late maturity before interruption by uplift. The Sulphur Mountain surface, cut during this cycle, had a probable relief of 2000 feet and was crossed by streams flowing in comparatively broad valleys. Sulphur Mountain stood higher than the undulating surface that truncated the folded Pliocene and Pleistocene strata of the Coastal Hills.

The Ventura River probably established its present course across the Sulphur Mountain Upland by headward erosion in Rincon shale at the east end of the Red Mountain dome. Near the close of the Sulphur Mountain cycle, Santa Paula Creek extended itself headward in the complexly faulted Pico and "Santa Margarita" rocks of the east-plunging anticline at the eastern end of Sulphur Mountain, and captured a west-flowing stream tributary to the Ventura River by way of Upper Ojai Valley.

Vertical uplift.—The Sulphur Mountain cycle was interrupted by vertical uplift of most of the Ventura region. The highest marine terrace on Rincon Mountain is 1300 feet, and the highest river terrace in the Ventura valley is 1170 feet. The river and marine terraces are warped, with the axis of uparching of the river terraces coinciding with the Red Mountain thrust.

Erosion, controlled by rock resistance and structure is responsible for the Sulphur Mountain escarpment, the form of Red Mountain, and the topographic expression of the larger faults. Most of the cliffs associated with faulting are resequent fault-line scarps, chiefly because the upthrown side of the larger high-angle reverse faults carries older, and generally more resistant, rocks above less resistant strata. San Cayetano thrust, an outstanding example, has thrust resistant Eocene sandstone over weak Pliocene and upper Miocene shale. Some cliffs are composite scarps, as Recent movement along the fault has produced a low fault scarp below the older and higher fault-line scarp. Such low fault scarps cut fans and terraces, as is well shown by the San Cayetano and Red Mountain thrusts.

The continuance of this type of deformation illustrates the essential continuity of the modern and past history of the Ventura region.

Correlation.—Several attempts (Gale, 1931, p. 60-76; Eaton, 1929, p. 713-762; 1932, p. 1-49) have been made to correlate events in the Cenozoic of the Ventura region with the standard glacial chronology of the eastern United States and western Europe. The chief difficulty in such a correlation is the absence of glacial deposits. A second, and almost equally serious handicap is the lack of agreement as to the duration of the Pleistocene and the relation of marine invertebrate fossils to the glacial sequence. Until this relationship is satisfactorily established, or more complete agreement is reached as to what constitutes the Pleistocene in California in terms of marine sedimentary rocks, it seems pointless to attempt a detailed correlation of events in this area with others several thousand miles away.

All that may safely be stated is that strata of the "Saugus" formation, early Pleistocene, were deposited in comparatively warm water following the accumulation of late Pliocene or early Pleistocene strata in cooler, and, more likely, deeper water. The depth of water may have had almost as great an ecologic control as climatic changes. After the deformation of these rocks in the principal orogeny, erosion produced a late mature terrane. At the close of the succeeding uplift the climate was slightly cooler than at present, as demonstrated by the Carpinteria fossils. The deposition of the Carpinteria gravels may possibly be synchronous with the Wisconsin glaciation of the central United States.

No attempt has been made by the writer to correlate terraces in the Ventura region with those of other localities in California (Fairbanks, 1897, p. 213-245; Smith, 1900, p. 179-230; Rode, 1930, p. 16-78; Woodring, 1935, p. 292-305). Such an attempt seems inadvisable in view of the nonuniformity of terrace elevations within the region,—obviously the last movement has been differential. The Oxnard Plain probably has been warped down almost as far as the terraced part of the coast has been elevated. Higher terraces of the Ventura region are the result of diastrophism rather than eustatic changes of sea level. For this reason, and the fact that the time of their cutting is necessarily restricted to the latest Pleistocene, it is not possible to correlate terraces of this region with glacial stages in the Pleistocene, as attempted by Davis (1933, p. 1041-1133) for the Santa Monica Mountains.

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SPITZENBERG CONGLOMERATE AS A TRIASSIC OUTLIER IN PENNSYLVANIA

BY LAWRENCE WHITCOMB

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ABSTRACT

The Spitzenberg conglomerate caps the conical hill of that name near Greenawald on the Hamburg, Pennsylvania, topographic map. No other outcrop of similar rock has been found in the surrounding region. Although actual contacts are buried, it seems that the conglomerate rests unconformably upon the Martinsburg shales, the conglomerate being in the form of a gently dipping spoon-shaped syncline, whereas the underlying Martinsburg is highly folded and has steep dips in all adjacent out-

crops. The lithology of the conglomerate is variable. Beds of arkosic sandstone, shale, and conglomerate are found, cross-bedding is common, most beds are gray, but some are reddish. The lithology is very similar to that of the Triassic fanglomerates of the Delaware Valley. No fossils have been found in the matrix or pebbles. The pebbles have been studied in an effort to determine their sources. While the Martinsburg and older formations seem to be lithologically incapable of furnishing the pebbles, numerous younger formations could have provided them. The location of the Spitzenberg in relation to the outcrops of these younger formations would require drainage from the north or west to bring the pebbles to their present location. Drainage in this direction is not believed to have existed in the Paleozoic, but it did exist in the Mesozoic. The fanglomerate type of lithology is common in the Triassic, but unknown in the Paleozoic of this part of the country. Although no paleontologic proof is available the Triassic age seems to be the most likely. If the Spitzenberg conglomerate is a Triassic outlier it becomes highly significant in the discussion of Triassic sedimentation in Pennsylvania.

INTRODUCTION

In recent years the study of the Triassic rocks in Pennsylvania and adjacent States has resulted in divergent opinions regarding the relationship between the drainage of Triassic and Recent times. As a result of this discussion, what was once a stratigraphic problem of only local interest now takes on new and greater significance. It is believed that the conglomerate to be considered in this paper may have an important bearing on this larger problem, and it therefore seems appropriate to reconsider the Spitzenberg conglomerate (Whitcomb and Engel, 1934) at this time.

LOCATION

The Spitzenberg conglomerate is found only as the cap rock of a single hill, the Spitzenberg, at Greenawald, 2½ miles to the northeast of Lenhartsville in Berks County, Pennsylvania (Hamburg quadrangle). (*See* Figure 1.) This hill, located within an area of Martinsburg shale (Ordovician), rises above the general level of the Harrisburg peneplain as a prominent landmark visible for a considerable distance.

As Lenhartsville is approached from the southwest on U. S. Route 22 from Hamburg the Spitzenberg stands out against the sky line as a distinct conical hill rising above the general level of the terrain (Pl. 1, fig. 1) to approximately 1100 feet above sea level. Since in this region the elevation of the Harrisburg peneplain is about 700 feet, the relief is about 400 feet. About halfway up the hill from the Harrisburg surface the slope steepens appreciably. This break in the profile is believed to represent the approximate location of the contact between the overlying conglomerate and the underlying shales and slates, although the actual contact has never been seen.

Five miles to the northwest is the main ridge of Blue Mountain which is composed of the Silurian Tuscarora (Shawangunk) conglomerate. Two and one-half miles to the west is "The Pinnacle", a spur produced

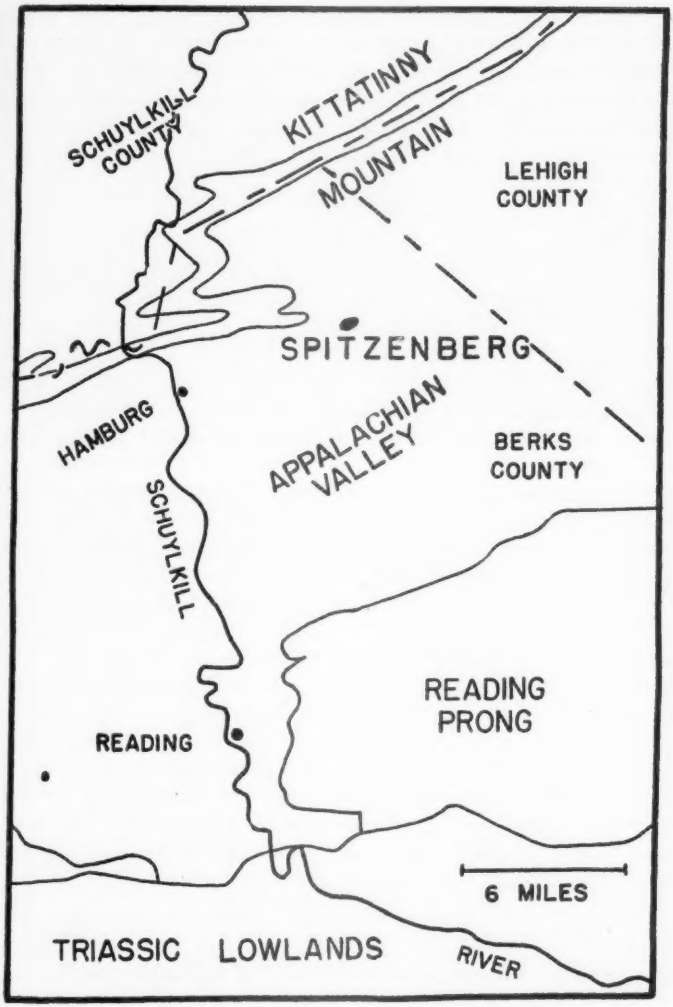

FIGURE 1.—*Outline map showing location of the Spitzenberg*

by an offset fold in the Blue Mountain; and one mile south of "The Pinnacle" is a secondary spur with an east-west axis, which is also formed of Tuscarora. The Spitzenberg lies to the south of the projection of the axis of this last fold and without apparent structural relation.

The elevation of the top of the Tuscarora ridges is about 1600 feet and represents the oldest peneplain surface preserved in the region. As previously stated the elevation of the Harrisburg peneplain is here approximately 700 feet, and therefore the Spitzenberg with an elevation of 1100 feet falls in between these two well-developed erosional surfaces and cannot be correlated with either of them.

GENERAL DESCRIPTION

LITHOLOGY

In any consideration of the Spitzenberg conglomerate a study of its lithology is very important. As one climbs the hill one first finds boulders of a coarse arkosic gray sandstone many of which have cavities from which limestone pebbles have been leached. These boulders rest on a soil which contains fragments of the underlying Martinsburg shale and the boulders are definitely brought down from higher ledges. At an elevation of approximately 900 feet, where the change in the slope of the hill becomes noticeable, are the first ledges of conglomerate and sandstone. From this point to the summit a rather complete composite section of the formation reveals constant and repeated change in the character of the sediments (Pl. 1, fig. 2). Sandstone and conglomerate beds of varying thicknesses are the rule. The sandstone beds show excellent cross-bedding and vary from gray to red. The arkosic character is noticeable. The conglomerate beds also have an arkosic matrix and contain pebbles of numerous lithologic types and sizes. The most common type of pebble is limestone, but there are also fragments of sandstone, quartzite, shale, and occasional pebbles of quartz and red jasper. The largest individuals in the conglomerate are the red sandstone cobbles some of which exceed 10 inches in greatest dimension (Pl. 2, fig. 1). The limestone pebbles seldom exceed 3 inches in size, the greatest percentage of them being smaller. There is no gradational sorting by size, and the axes of longest dimension are at varying angles to the bedding. Lenses of sandstone or conglomerate may be seen (Pl. 2, fig. 2).

CONDITION OF DEPOSITION

Study of the many exposures strongly suggests that the deposit was laid down as an alluvial fan by a stream that was subjected to repeated and marked changes in velocity. Examination of the fanglomerates of known Triassic age in New Jersey and Pennsylvania has shown that the type of lithology is definitely comparable.

STRUCTURE

As previously stated, it is not possible at any point to observe the actual contact of the conglomerate and the underlying Martinsburg

FIGURE 1. SPITZENBERG RISING ABOVE LEVEL OF HARRISBURG PENEPLAIN
Taken from a point 3 miles to the southwest. Telephoto X3.

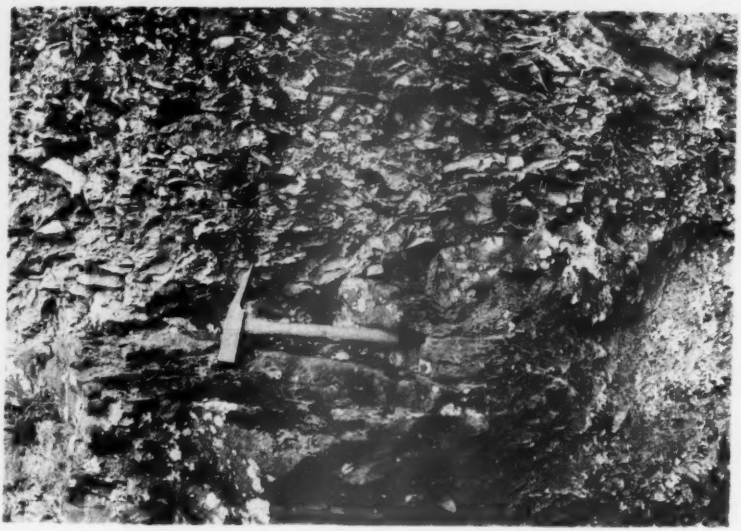

FIGURE 2. CONGLOMERATE ABOVE LENS OF REDDISH SANDSTONE

SPITZENBERG CONGLOMERATE

FIGURE 1. TEN-INCH RED SANDSTONE COBBLE IN CONGLOMERATE

FIGURE 2. IRREGULAR LENSES OF SANDSTONE AND CONGLOMERATE
Height of area in view approximately 3½ feet.

shale. Its approximate location may be determined by the fact that the shale float is found on the slopes of the hill up to the place where the topographic break is noticed and that shortly above this elevation one finds the first outcrops of the conglomerate. Accordingly, the contact should be at about 900 feet which would give a thickness of approximately 200 feet for the conglomerate. Accurate measurement of the strike and dip in the different outcrops is hindered by the lensing and cross-bedding. However, all measurements made at varying elevations and on different sides of the hill indicate that the conglomerate is a gently dipping spoon-shaped syncline. In sharp contrast, examination of the Martinsburg shale and slate in the many outcrops of the surrounding region shows that it has been highly folded, steeply pitching isoclinal folds being seen in some places, and that slaty cleavage has been developed in the lower slopes of the hill. An unproven assumption would be that there is a marked angular unconformity between the two formations.

DETERMINATION OF AGE

PREVIOUS WORKERS

This conglomerate has been mentioned by previous workers, but there has been no definite statement as to its age. Behre (1933, p. 142-143) has suggested that it may represent a phase of Martinsburg sedimentation but also mentioned that it might be of Triassic age. Stose (1930, p. 650-651) writes of the synclinal structure but leaves the age as undetermined, although he now leans to the Triassic age (oral communication, 1940). Whitcomb and Engel (1934) expressed their belief that the most probable age was Triassic.

PALEONTOLOGIC EVIDENCE

Repeated attempts have been made to find fossils. Thousands of pebbles have been examined in the hope that a fossil might be found which would at least give some limiting age to the conglomerate, but to date not a single one has been found in the pebbles or in the matrix.

POSSIBLE PALEOZOIC AGE

General statement.—As there is no paleontologic evidence, other methods must be used to determine the stratigraphic position of this formation. Certainly it is younger than the Martinsburg shales on which it rests; the idea of a thrust sheet of older rocks is too fantastic. If the conglomerate is Paleozoic three alternatives require consideration: (1) It is middle or upper Martinsburg; (2) it is an eastern equivalent of the Bald Eagle of central Pennsylvania; or (3) it is Tuscarora.

Martinsburg.—In his report Behre (1933, p. 142-143) "for want of a better interpretation" places the Spitzenberg conglomerate in the middle

Martinsburg but says that it is placed there "with serious reservations." Other people who have seen the conglomerate have suggested that it was upper Martinsburg, a difference of opinion largely controlled by one's ideas of Martinsburg structures in the region. There are some very serious objections to this Martinsburg age. One of the greatest difficulties is the source for the various types of pebbles in the conglomerate. By middle or late Martinsburg time the older rocks had been pretty well buried under Martinsburg sediments. Today the nearest outcrop of the Cambrian and Ordovician limestones is 6 miles away; at the time of conglomerate deposition it must have been farther from a possible source of Cambrian and Ordovician limestone pebbles. Limestone pebbles could have been derived from some of the thin limestone beds in the Martinsburg if they had been exposed by folding and erosion, but it is doubtful if such a great quantity could have been produced without a great number of definite Martinsburg shale and slate pebbles which are not found. Another difficulty in assigning a Cambrian or Ordovician age to the limestone pebbles is that lithologically they do not seem to be like most of the limestones of that age. Chemical analyses of typical limestone pebbles (Whitcomb and Engel, 1934, p. 40) do not agree with the results that are obtained from the Cambrian limestone of the Lehigh Valley and would correspond to only certain beds in the Beekmantown. Furthermore there is no known red sandstone of pre-Martinsburg age that could have produced the pebbles and cobbles of this lithologic type. To be sure there are some red sandstone beds in the Martinsburg itself, but the lithology of these beds does not seem to be comparable to that of the cobbles. Another great difficulty is to account for the type of sedimentation represented in the Spitzenberg. Our present ideas regarding the conditions of sedimentation in Martinsburg time do not explain how such a conglomerate could have been formed well within the area of Martinsburg deposition. Such a deposit should have been laid down near a highland by a stream which in its upper reaches flowed over rocks of varied types. No such highland is believed to have existed in this region during Martinsburg time. It is therefore believed that the Martinsburg age cannot be accepted.

Bald Eagle.—It has been suggested that the Spitzenberg conglomerate might be equivalent to the Bald Eagle conglomerate of central Pennsylvania. On the west bank of the Susquehanna River in the water gap through Blue Mountain is about 30 feet of gray sandstone and conglomerate which Willard and Cleaves have assigned to the "Oswego" (1938, p. 6-7) or the eastern equivalent of the Bald Eagle. At the Lehigh Gap the first 2 or 3 feet of conglomerate which directly overlies the Martins-

burg has also been tentatively assigned to the Bald Eagle by Willard and Cleaves (1939, p. 1184-1185). If this determination at the Lehigh Gap is correct the Bald Eagle was deposited after Taconic folding. On the basis of this evidence it might be suggested that the Spitzenberg conglomerate was of the same age and that it was deposited after the completion of the Taconic disturbance which would possibly explain the type of sedimentation. Such an age, however, would not overcome the difficulty of finding a source for the pebbles. Another difficulty with this proposed age is in trying to match the lithology of the two formations, for wherever known in this part of the State the Bald Eagle weathers to a rusty surface, which is not the case with the Spitzenberg conglomerate. Again the types of pebbles in the two conglomerates are distinctly different, the Bald Eagle being essentially quartz and quartzite. It is not believed that Bald Eagle age can be established.

Tuscarora.—As the Tuscarora is the main "ridge maker" in the region it must be considered in any discussion of the Spitzenberg conglomerate. Lithologically the two formations are distinctly different as the pebbles in the Tuscarora are largely white quartz, and the limestone pebbles which are so abundant in the Spitzenberg conglomerate are practically nonexistent. Of the many geologists who have examined the Spitzenberg conglomerate not one is known who would seriously correlate it with the Tuscarora.

Upper Paleozoic.—Although several people in searching for a solution to the age of the Spitzenberg conglomerate have considered some of the higher conglomerates in the Paleozoic, none has attempted to place the formation in the higher portion of the series. Too many structural, depositional, and lithologic problems obstruct such a correlation.

PROBABLE TRIASSIC AGE

Lithology.—Difficulties disappear if the Triassic age is accepted. In the first place it is a well-established fact that fanglomerates were formed in Pennsylvania and New Jersey during the Triassic. The lithology of the Spitzenberg conglomerate most certainly resembles more closely these known Triassic fanglomerates than any of the known Paleozoic deposits. Although the individual pebbles and cobbles are not of the same lithologic types as those found in some of the Triassic fanglomerates along the Delaware River, they are not expected to agree perfectly, for each fanglomerate would be formed by streams draining from areas of different lithology. In a case of this type the conditions of sedimentation would be more significant than the character of the individual component parts. Secondly, the problem of the source of the pebbles no longer causes any

trouble, for under the conditions of Triassic sedimentation the streams would have been flowing from the northwest toward the southeast across highly variable folded sedimentary rocks. Under such conditions the red sandstone cobbles would have a source in similar beds of the upper Paleozoic; and the limestone pebbles difficult to account for under Paleozoic conditions also would be easily obtained from Silurian or Devonian limestones that would have been exposed to erosion to the northwest. Thirdly, the question of how a deposit which shows the characteristics of having been laid down by torrential streams could have been formed well within the area of Martinsburg sedimentation and at such a distance from any known highlands is no longer a stumbling block, for the newly formed Appalachian Mountains would provide the highland. Furthermore the fanglomerates would form near these mountains, which is the case in all other Triassic fanglomerates.

Objections have been raised as to the suggested Triassic age, but it is not believed that they are valid. It has been pointed out that no known pebbles can be definitely correlated with the Tuscarora over which any stream from the northwest must have flowed. This is not considered important, for in the Triassic fanglomerates in the Delaware River Valley where the stream had come over the pre-Cambrian rocks the fanglomerates show no abundance of pre-Cambrian pebbles. This lack of pebbles of the last hard formation crossed by the stream is caused, it is believed, by the fact that a river gathers its load of sediment in the upper portions of the stream and that it would have very little active erosional power as it crosses the last hard formation before deposition takes place.

Structure.—Some have argued that the supposed angular unconformity between the Martinsburg and the overlying Spitzenberg conglomerate could have been produced if the conglomerate were a competent bed of the Martinsburg resting upon incompetent beds. Admitting that such structures could be produced, it does not remove the difficulties encountered in obtaining the pebbles of varied lithologic types nor does it account for the conditions of deposition that are needed to produce the deposit.

Geographic location.—Another problem concerns the position of the Spitzenberg conglomerate in relation to the other known Triassic fanglomerates. Lying in a region of Paleozoic rocks some 18 miles to the north of the nearest known Triassic, it has been stated that it cannot be Triassic. Those who put forth this argument believe that the Triassic fanglomerates were deposited along the line of a series of active faults which bound the western edge of the Triassic basins and that therefore

its Triassic age is untenable. Others (Wherry 1913, p. 118; Bascom, et al., 1931, p. 47) believe that the western boundary of the Triassic is in some places a normal sedimentary overlap upon the older rocks and not always a fault contact. If this should prove to be the case this argument against the Triassic age cannot be sustained.

Triassic drainage.—The problem of the elevation of this conglomerate in relation to that of the various peneplains of the region is very important for it raises the question of the relationship between the drainage of Triassic and Recent times. If one accepts the idea that the streams of Pennsylvania were superposed upon the arched Fall Zone peneplain and its Coastal Plain cover as suggested by Johnson (1931), difficulties are immediately encountered. According to this hypothesis it would be impossible to get Triassic sediments in the position now occupied by the Spitzenberg conglomerate without major faulting, of which there is not the least evidence in this particular area, for the horizon at which the conglomerate is found is below the level of the Fall Zone and Schooley peneplains neither of which was supposed to have been developed until after Triassic time. This would be a serious argument against the Triassic age of this formation if the superposed origin of the drainage were proven. According to Meyerhoff and Olmsted (1936) the idea of superposed drainage has not been proved, and they believe that certain of the present streams, including the present-day Schuylkill, "were well established before Newark displacement and deposition began, and . . . maintained their courses throughout the entire epoch of deformation" (1936, p. 38) and further that the present streams are still flowing in essentially the same valleys. If this concept of the origin of Appalachian drainage is considered it becomes obvious that under such a set of conditions it would be possible to have the formation of a Triassic fanglomerate along the course of one of the tributaries of the Schuylkill. Under such conditions the problem of the elevation of the conglomerate with regard to the height of the peneplains can be explained and is not a major obstacle.

CONCLUSIONS

It is admitted that the Triassic age of the Spitzenberg conglomerate has not been proved beyond a doubt but it is believed that a strong case has been made. If the Spitzenberg conglomerate is of Triassic age it becomes of vital importance in the solution of the problem in Appalachian drainage for the sequence of events in the development of the superposed drainage of the Appalachians as proposed by Johnson (1931) will have to be reconsidered.

Some may believe that undue importance is being placed upon the age of this particular conglomerate, but any hypothesis like that of superposed Appalachian drainage must be examined with great care before it is accepted as a fact. It is important to remember that mere number of arguments in favor of a hypothesis are of little value as proof if a major difficulty cannot be overcome. The proven Triassic age of the Spitzenberg conglomerate would be a major difficulty to the idea of superposed drainage. It is hoped that increased interest in this conglomerate will eventually lead to conclusive proof of its age and thereby to a more complete understanding of the development of Appalachian drainage.

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SPECTROSCOPIC DISTRIBUTION OF MINOR ELEMENTS IN IGNEOUS ROCKS FROM JAMESTOWN, COLORADO

BY JOSEPH M. BRAY

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ABSTRACT

Spectrographic analyses of a group of igneous rocks and their constituent minerals from Jamestown, Boulder County, Colorado, revealed:

Qualitative Analyses

Rocks: Each, regardless of age, is distinguishable by variations in content of certain elements (Sc, La, Ce, Nd, Zr, V, Cr, Co, Ni, Pb, and Zn). Comparison with several Massachusetts rocks emphasizes the province relationship.

Minerals: Minor elements occur systematically. Sr is highest in K and Ca minerals (particularly feldspars); Ba in K minerals (potash feldspar and hornblende). All quartz and sphenes contain Al. Sc is most abundant in Mg and ferrous Fe minerals (micas and hornblende). Sphenes contain a remarkable proportion of rare earths. Ti, Mn, V, and Cr are concentrated in the dark minerals and muscovite. Ni and Co occur almost exclusively in biotite.

Quantitative Analyses

Quantitative analyses substantiate all conclusions based on the qualitative work.

Minerals: Characteristic minor elements of each mineral vary from rock to rock. The SrO : BaO ratios decrease numerically in the order of plagioclases, biotites, K-feldspars, and muscovites. Pegmatite K-feldspars have much larger SrO : BaO ratios than parent rock samples. This is explained by temperature-crystal structure relations. Qualitatively and quantitatively each pegmatite mineral bears an average of less of almost every minor constituent than the same mineral in the parent rock. This also is explained by temperature-crystal structure relations.

The minor constituents are probably present in solid solution.

INTRODUCTION

NATURE AND PURPOSE OF INVESTIGATION

Most chemical analyses of igneous rocks report only the abundant elements. The less common ones have been studied, particularly in Europe, but almost exclusively from the broad viewpoint of geochemical

distribution. They are important from the purely petrological standpoint, however, and aid in deciphering the distribution and concentration of many metals in ore deposits. The role of pegmatites in the evolution of igneous rocks is clarified by such studies, as is the problem of hydrothermal origin of certain igneous-appearing rocks.

The following fundamental questions confront the investigator: Are ore deposits related only to igneous rocks of purely magmatic origin, or do they form in association with igneous-appearing rocks of hydrothermal origin? Can ores and pegmatites be derived from the same igneous mass and occur at different depth levels governed by pressure alone? Are metallic elements in an igneous mass incorporated as minor constituents in rock-forming minerals if not tapped off at the critical time to allow migration into veins? What is the role of pegmatites in the concentration of minor elements? It was with these questions in mind that this work was undertaken. Much more detailed research is necessary before the final answers can be given.

Before this detailed study was undertaken, a suitable petrographic province had to be selected, which possessed igneous rocks of diverse types and ages, and if possible associated with both ore deposits and pegmatites. Competent, detailed mapping was a prime prerequisite. The Jamestown district, Boulder County, Colorado, met these requirements, so the author collected 140 representative specimens.

The original plan was to investigate minor elements in rock samples only, but since the individual minerals were found to hold the key to the situation particular emphasis was placed on their study.

The research was carried out in the Cabot Spectrographic Laboratory of the Department of Geology at the Massachusetts Institute of Technology, Cambridge, Massachusetts. Both qualitative and quantitative analyses were made.

In this study conclusions were fitted to the data. The author had no preconceived ideas. No data were omitted because of disagreement with the conclusions drawn.

ACKNOWLEDGMENTS

The writer wishes to express his gratitude to Professor Walter H. Newhouse, of the Massachusetts Institute of Technology, who suggested the problem and gave helpful criticism and advice. Much credit is due Mr. R. F. Jarrell, who guided the writer in practical spectrography, conferred on innumerable problems in technique, and critically read portions of the manuscript. To Professor T. S. Lovering, of the University of Michigan, thanks are extended for his interest in the problem.

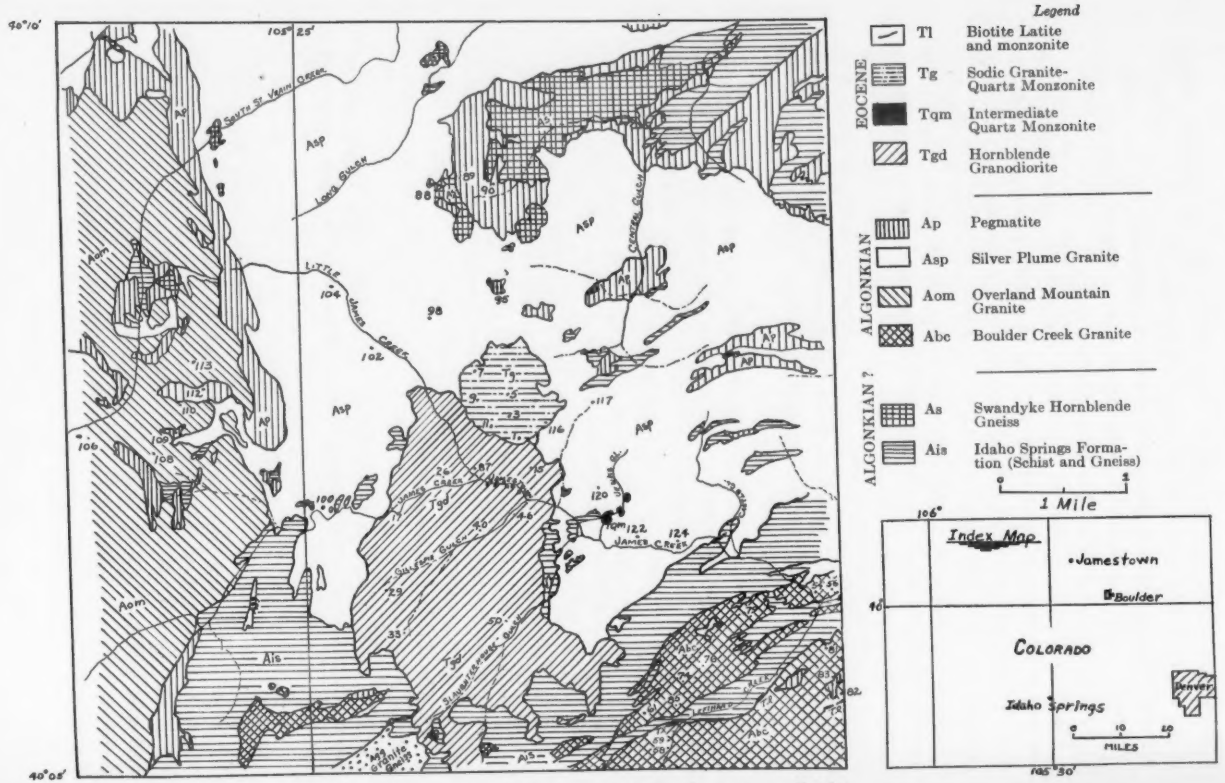

FIGURE 1.—*Geologic map of the Jamestown district*

his co-operation and aid in the selection of a suitable geological province, and for furnishing a number of specimens. Professor M. J. Buerger, of the Massachusetts Institute of Technology, gave valuable counsel in matters pertaining to crystal structure. Mr. Albert J. Jehle, of the Massachusetts Institute of Technology, deserves special thanks for his tireless laboratory assistance. Mr. Charles E. Hicks, of Lafayette College, aided in the final drafting. The consultation of Dr. E. N. Goddard, of the United States Geological Survey, was of great help.

GEOLOGY OF THE DISTRICT

GENERAL

Jamestown is located in central Boulder County, Colorado, 10 miles northwest of Boulder, 35 miles northwest of Denver, at the northeast extremity of the Front Range mineral belt. Figure 1 is a generalized geologic map of the district, based on the work of Lovering and Goddard (1939). The rocks are pre-Cambrian schists and gneisses, intruded by middle pre-Cambrian Pikes Peak and Silver Plume granites (details of the Front Range geology are given by Ball, 1908, Boos and Boos, 1934, and Lovering, 1929) and, during the Laramide Revolution in late Cretaceous or early Eocene time, by a series of stocks and dikes ranging from diabase to alaskite. The Laramide igneous sequence has been discussed by Lovering and Goddard (1938a; 1938b).

PRE-CAMBRIAN ROCKS

The pre-Cambrian rocks, from oldest to youngest, are as follows:

(1) The Idaho Springs formation, a metamorphosed sedimentary series, consists of varied schists and injection gneiss genetically related to the Boulder Creek granite intrusion.

(2) The Swandyke hornblende gneiss was not considered in this work.

(3) The Boulder Creek granite, the older member of the Pikes Peak group, is highly variable in texture and mineral proportions, because some is intrusive granite, some granitized schist, and some granite containing large quantities of assimilated schist. It is a medium- to coarse-grained, gray biotite-rich quartz monzonite. Microcline and plagioclase constitute 50 per cent, and biotite and muscovite 10-30 per cent of the rock. Quartz is abundant. The accessories are magnetite, apatite, zircon, and rarely rutile (?). Epidote and zoisite were seen in one thin section. Minor sericite and carbonate are alteration products. A coarse-grained facies of this rock in the southern part of the district along Lefthand Creek contains pink microcline phenocrysts (with long axes parallel) up to an inch long.

FIGURE 1.—*Geologic map of the Jamestown district*

(4) The Overland Mountain granite, classed by Lovering and Goddard (1939) as Boulder Creek granite, is here given a local name, following Goddard's original idea (U. S. Geol. Survey, unpublished report). It is the younger member of the Pikes Peak group. The gneissic border closely resembles Boulder Creek granite. The rock, very coarse-grained, feldspathic, and containing microcline phenocrysts in a coarse groundmass of andesine, quartz, and biotite, is like a fine-grained pegmatite (which contains coarser pegmatites). The accessories are magnetite, apatite, sphene, and zircon. Quartz-feldspar intergrowths are present. The quartz frequently includes rutile needles.

(5) The Silver Plume granite, youngest of the pre-Cambrian series, underlies most of the district. It is medium-to coarse-grained, frequently porphyritic. Flow structure is often well developed. The phenocrysts are orthoclase, and the groundmass quartz and oligoclase, with some mica. Accessories are magnetite, apatite, and rare rutile. Biotite often exhibits well-developed haloes surrounding zircon inclusions.

(6) Each pre-Cambrian granite has related dikes and zones of pegmatite. The chief minerals are microcline, quartz, and muscovite. Graphic textures are common in the pegmatites of the Silver Plume granite. Cerite deposits occur locally in narrow aplite-pegmatite seams related to the Silver Plume (Goddard and Glass, 1940).

TERTIARY ROCKS

The Tertiary igneous sequence is as follows (Lovering and Goddard, 1938a), from oldest to youngest:

(1) Diabase dikes are confined to the extreme southwest portion of the district.

(2) The hornblende granodiorite group forms a large stock in the south-central part of the area and a series of dikes to the west on Overland Mountain. The stock is chiefly medium-grained granodiorite. The main minerals are hornblende, labradorite-andesine, orthoclase, and lesser quartz and magnetite. Hornblende is very conspicuous. The texture is even-grained and dense. Sphene is an important accessory, and apatite is also present. Epidote and chlorite are alteration products. Biotite and pyroxene are present near the borders of the stock.

(3) The intermediate quartz monzonite and (4) the alaskite porphyry groups will not be considered.

(5) The granite-quartz monzonite group, forming a stock and series of radial dikes in the central part of the district, is closely related to the lead-silver and pyritic gold ores. The stock is compound, a quartz monzonite to the northwest and a granite to the southeast. Orthoclase

phenocrysts are abundant in the granite. The groundmass contains smaller phenocrysts of albite-oligoclase with orthoclase, quartz, biotite, and magnetite. Accessories are apatite, sphene, and zircon. The entire stock has been altered, and feldspars are often kaolinized or sericitized.

(6) Bostonite porphyry dikes will not be considered.

(7) Biotite monzonite and latite dikes, thought to be genetically related to the telluride gold ores throughout the Front Range mineral belt, will be considered by the writer in a separate paper.

The ores of the Jamestown district have been fully discussed by Goddard (1935).

PREPARATION OF SAMPLES

PRELIMINARY

Preliminary field excursions and study of maps showed the necessity of sampling at frequent intervals. Special attention was paid to freshness and exact location. Numbers on Figure 1 locate the samples. Pegmatite samples were collected only from within each parent granite mass, in order to avoid confusion concerning the source rock. New paper bags were used to prevent contamination. Thin sections were studied.

CRUSHING AND SEPARATION

Samples were split, washed, dried, and crushed on an anvil of the purest steel obtainable (it contained 0.25 per cent of Mn, 0.20 per cent Si, and 0.025 per cent P and S).

Samples for qualitative analysis were crushed, quartered, and powdered in a clean agate mortar. Individual mineral samples were prepared by breaking the rock to a medium-fine size, washing to remove dust, and picking by hand under the microscope. The particles readily adhered to a pointed, moist toothpick. The minerals were distinguished by ordinary physical properties. Magnetic concentrates contained some biotite and hornblende which were eliminated by finer grinding and retreatment with a weak magnet (the magnetic concentrates never came into contact with the magnets).

Some precautions taken to prevent contamination in preparing the samples were as follows: (1) Glass vials in which the final samples were kept were thoroughly cleansed in hot concentrated nitric acid and distilled water, (2) tools used were washed and cleaned between samples, (3) mineral concentrates were repicked to eliminate grains of other minerals, and (4) nonmagnetic samples were placed beneath a strong electromagnet to remove possible adhering particles of steel or iron minerals.

SEPARATION OF POTASH FELDSPARS FROM PLAGIOCLASES

The technique described by Gabriel and Cox (1929) and modified by Keith (1939) was followed. No contamination resulted from the staining procedure, since no cobalt appeared in the feldspar analyses.

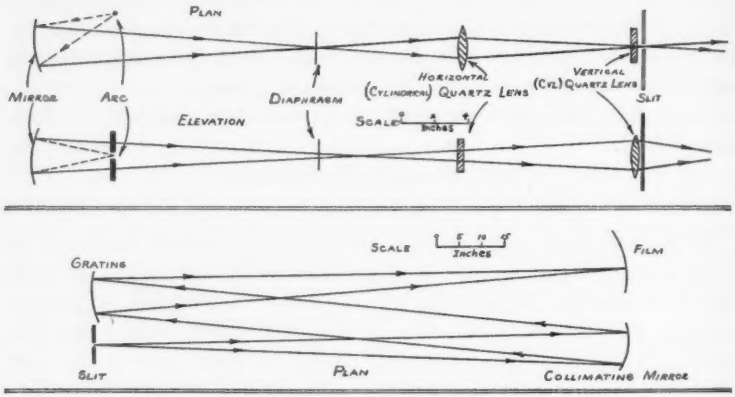

FIGURE 2.—*Drawing of spectrograph system*
Upper: arc to slit. Lower: slit to film. (Wadsworth mounting)

ADDING INTERNAL STANDARD

All samples were powdered, and 30 milligrams were mixed for 10 minutes with 10 milligrams of internal standard mixture in an agate mortar. The mortar was soaked in hot concentrated nitric acid between crushings and mixings.

ANALYTICAL METHODS

GENERAL

The instrument used was a 21-foot, 30,000-line (per inch concave) grating spectrograph on a Wadsworth mounting. Figure 2 illustrates the set-up. The useful range of the instrument was from 1800 Å to 5600 Å, and the dispersion at the setting used was 2.47 Å per millimeter over the entire film.

Preliminary experimental work resulted in choice of the cathode layer method of carbon arc spectral analysis. The method has been described by Strock (1936b). All common spectroscopic methods depend upon vaporization of a sample and recording of the light emitted. In the cathode layer method the sample is vaporized from the cathode. Two advantages of the method are its great sensitivity (one part in a million often determined in this work) and the fact that it requires only a very

small sample. The cyanogen background on the photographic plates is denser for the cathode layer than for other methods but can be overcome. A spectrograph of high dispersion is necessary to separate the numerous lines on the cathode layer spectrum, and great care is necessary to insure uniformity of analytical procedure.

OPERATING TECHNIQUE

Figure 2 shows that the arc radiation (produced by electrical excitation) is directed into the spectrographic system by an aluminized mirror displaced slightly from the optical axis of the instrument. From the mirror the radiation is focused on a diaphragm, beyond which it passes through a horizontal and a vertical quartz lens, then through the slit. The light passing the slit reaches a collimating mirror, then the grating, and then passes to the photographic plate. A complete record of the light radiation, and therefore of the elements present, is made.

The electrodes were National Carbon Company specially pure $\frac{3}{16}$-inch carbon (cathodes) and graphite (anodes). The cathodes were 20 millimeters long and had a 2-millimeter hole drilled in one end to a depth of 6 millimeters. The outer edge of this end was turned down 5 millimeters to form a cup in which the sample was placed. Eastman Safety Commercial film was used. Films were developed in a special solution for exactly 5 minutes at 18° C, fixed in a special hypo solution for 7.5 minutes, and washed for at least an hour.

QUALITATIVE ANALYSIS

Preliminary qualitative analyses detected the minor elements present. The elements chosen for study were those which showed significant variations from mineral to mineral or from rock to rock. Almost all these elements had sensitive spectral lines between 2800 Å and 4650 Å, the range (covered by one photographic film) chosen. It must be emphasized that the spectral range used does not contain sensitive lines of all elements that might have been worth studying. For instance, sensitive lines for Li and B were not in this range. In addition, lines for a number of desirable elements (such as Nb, Ga, Rb, and Cs) were either too diffuse, too near sensitive lines of a major element, or obscured by cyanogen lines. The rare earths not listed in the analyses were not determined because of confusion with other spectral lines. Some electrodes contained small quantities of Cu, so that element was disregarded.

The spectra were critically studied by visual comparison with a master plate bearing lines of most known elements. The most sensitive lines for the elements were used wherever possible. A semiquantitative aspect was given to the qualitative results by adoption of an arbitrary scale based on the variations of an iron line reduced in steps by a step sector,

expressing the relations between quantities of elements present and densities of their spectral lines. The scale is as follows:

Symbol		Quantity	Density of line	
A	—	Absent	(line absent)	
VT	—	Very small trace	(very light line)	(Variations within one density range are expressed by plus and minus signs)
T	—	Trace	(light line)	
S	—	Small	(distinct line)	
M	—	Medium	(medium line)	
L	—	Large	(heavy line)	
VL	—	Very large	(very heavy line)	
EX	—	Extreme	(heaviest line)	

No absolute percentage figures can be assigned to these steps, and symbols should be used only in comparisons involving a single element. The sensitivities of various elements vary widely, so that a quantity designated as "large" for one might really represent less than that indicated by "small" for another.

Samples of 48 rocks and 69 samples of individual minerals were analyzed qualitatively. The procedure used for the rocks differed from that used for the minerals, so the results for the two groups should not be compared. Larger samples were used for rocks, and the current was applied in increasing increments at intervals of 15 seconds. In the work on minerals less sample was used, and the current was kept constant. The lines of the abundant elements Na, K, Si, Al, Fe, and Ca masked no lines of the minor elements.

QUANTITATIVE ANALYSIS

Quantitative analyses were restricted to mineral samples, each made in duplicate, so figures given are averages of two determinations. The internal standard method, refined by use of iron step-spectrum calibration of each plate and photometric density measurements, was used. An iron-calibration step spectrum was photographed on each film, together with the unknown spectra. Jarrell (paper in preparation) will discuss the theory of quantitative spectral analysis as applied to geological problems, so theory will be largely omitted here. Since each problem is a special case of method, however, brief mention of methods is desirable.

All samples were 2-4 milligrams. Electrode separation was maintained at 5 millimeters, and slit width at 0.05 millimeter. During calibration with the iron arc the current was kept at 3.5 amperes, and the potential at 35 volts. All samples were vaporized at 7.5 amperes and 65 volts. The same film, same strength developer and fixer, and the same time and temperature of development were maintained throughout.

The step sector used for calibration (using the iron arc) was cut so that the relative exposure of each successive step increased as a geometrical progression of factor two. The sector was spun at high speed

and placed as close to the slit as possible. The density decrease of each step in the iron spectrum is proportional to the logarithm of the exposure (log. E). The exposure factor is the result of a number of variables.

After proper setting of the microphotometer galvanometer deflection on a horizontal scale for no light and for the uninterrupted light beam, the photometric density of any spectral line was read directly as the galvanometer deflection on the horizontal scale when that line was moved across the light beam. For each step in the spectrum, the negative logarithm of the line's deflection (in millimeters) was plotted simply by plotting the density readings on semilogarithmic paper. The deflections for each step were plotted as ordinates against the logarithm of the corresponding sector exposures (1, 2, 3, 4 seconds), representing the log. E values, as abscissae. The curves are linear except in their upper portions, where they approach the abscissa asymptotically (Fig. 3). These curves were the basis for calibrating each film. It was necessary to have different iron calibration curves for lines separated by more than 25Å—100Å, depending on the spectral region. Thus there was a group of calibration curves for each film. The line used for each element was referred to the same iron calibration curve on every film. Figure 3 is a sample calibration curve, actually used for the Na and Pd lines on one film. For such curves the negative logarithm of exposure (log. E) for the line of an unknown element can be read from the abscissa, after the line deflection is obtained with the photometer. For positive quantitative results, the density of the line of each element present in unknown quantity must be compared to a suitable line of some element always present in known concentration, called an internal standard. This comparison was made as follows: The negative log. E values for both internal standard and unknown were obtained, in each case using the galvanometer deflection and proper iron calibration curve. The negative log. E for the unknown was subtracted from that for the internal standard. The difference was called the Δ log. E, or difference in log. exposures. For each element in each mineral a curve had already been prepared, plotting Δ log. E values against known concentrations of that element in a prepared mineral sample. Therefore, once the Δ log. E value for any unknown spectral line was obtained, the exact concentration of the corresponding element in the sample was read from the standard curves. The preparation of these standard curves will be discussed later.

The lines of a number of important elements were too dense on the films to be read with the photometer (zero deflections). A "reducing" sector was used to cut the density of these lines by limiting the light passing to the film. The reducing sector, consisting of two adjustable

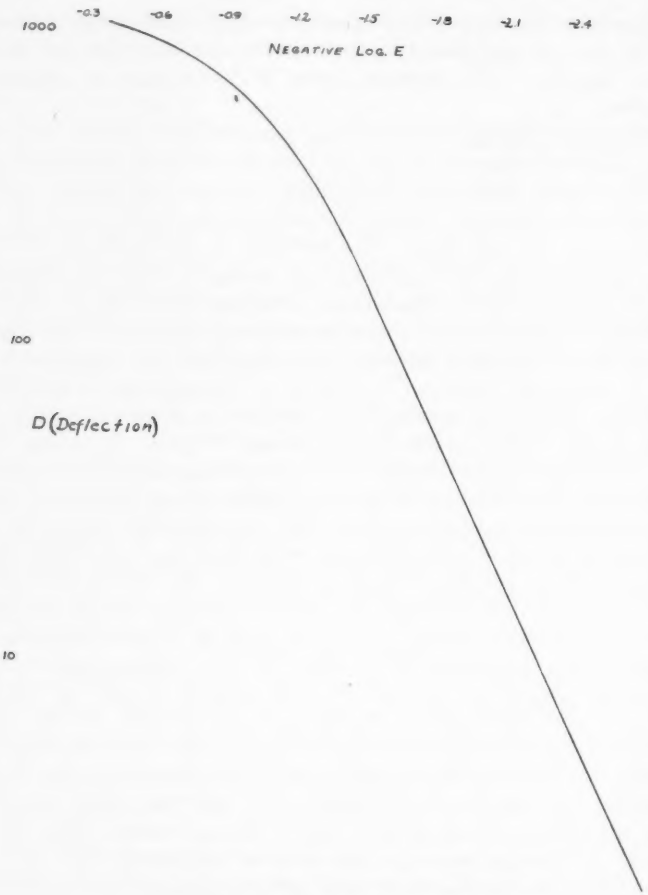

FIGURE 3.—*Sample calibration curve*

(Fe 3279.15 A) used for Cu, Na, and Pd lines on film XXXIV. Abscissa: log. relative exposure (1, 2, 4, 8, 16, 32, 64 seconds). Ordinate: actual density measured (deflection)

discs, was set for a reduction of 32:1. A double spectrum was obtained by setting the sector to cover only half the slit. Half of the spectrum was of normal density, the other half much lighter. The lines too dense on the normal spectrum were read on the lightened part. The iron calibration lines and those of the internal standard (Pd) were, of course, lightened in the same proportion as those of the unknown elements.

Each calibration curve must lie in exactly the same position on the plotted diagram relative to the internal standard calibration curve for every film before comparisons from film to film are permissible. This was not possible, because of conditions beyond control, so a base film

TABLE 1.—*Spectral lines used in the quantitative work*

Material obtained from Harrison and staff (1939).

Element	Line used	Arc intensity	Sensitivity	Internal standard line	Iron calibration
Mg I	2852.129 Å	300 R	U 1		2852.1
Ni I	3101.554	1000 R	—		3180.23
Pd I	3242.703	2000 Wh	—	3242.703 Å	
Cu I	3247.540	5000 R	U 1		3279.15
Na I	3302.988	300 R	—		
Zn I	3345.572	500	—		
Ti II	3349.406	100	—		
Ni I	3414.765	1000 R	U 1	3421.240	3415.534
Pd I	3421.240	2000 R	U 2		
Co I	3453.505	3000 R	—		
Pd I	3481.152	500 R	—		
Mn I	4034.490	250 r	U 3		4073.45
Ca I	4226.728	500 R	U 1		
Sc II	4246.829	80	—		
Cr I	4254.346	5000 R	U 1		
Fe I	4271.764	1000	—	3481.152	4246.09
La II	4333.734	800	—		
V I	4379.238	200 R	U 1		
Ba II	4554.042	1000 R	V 1		
Sr I	4607.331	1000 R	U 1		4547.85

I—Line classified as emitted by normal atom.
II—Line classified as emitted by singly ionized atom.
R—Wide self-reversal.
Wh—Wide, hazy.
r—Narrow self-reversal.
For neutral atoms the most sensitive line is designated as U 1, and less sensitive lines in order as U 2, and so on. For the singly ionized atom the designations are V 1, V 2, and so on.
The figures in the intensity column are arbitrary, based on the intensity for the pure element.

was selected, to which all others were referred upon application of appropriate corrections. Before analytical figures were recorded the calibration curve for each element on every plate was corrected so that it was the same horizontal distance from the appropriate internal standard calibration curve. Table 1 lists the spectral lines used, together with other pertinent data.

INTERNAL STANDARD

Since different minerals were analyzed, and no element could be present in all in constant proportions, a known quantity of some element had to be added. After due experimentation a 3% mixture of palladium as palladium chloride in carbon was chosen as the internal standard, because of its purity, similarity between its boiling point and those of

the elements to be determined, and because it had lines in several parts of the spectrum.

The internal standard element should have lines near those of the elements to be determined, so that both can be referred to the same calibration curve. No useful palladium lines occurred above a wave length of 3481.15 Å (when the proper concentration was used), so that all lines of unknown elements above this wave length were referred to iron calibration lines in their immediate vicinities. The Pd internal standard line at 3481.15 Å was referred to the iron calibration line at 3415.53. A comparison was then made between the internal standard and unknown lines according to the method outlined previously. This method was proven satisfactory by experimentation. All Pd lines on all plates bore a constant relation to one another. The same procedure was used in preparing the standard concentration curves.

It must be remembered that a large number of elements were determined by a single procedure. Under such circumstances exact control and regulation of methods and conditions are more difficult than if only one element is to be determined, such as many investigations carried out by Goldschmidt and his coworkers.

The most effective proportion of mineral sample to internal standard base was 30 milligrams to 10 milligrams. Three separate internal standard bases were prepared, one each for quartz, mica, and feldspar.

STANDARD MIXTURES

Mixtures containing known concentrations of the elements to be determined were prepared, as follows:

Element	Source	Quantity (mg.)
Ba	BaO	50
Sr	SrO	50
Ti	TiO_2	50
Cr	Cr_2O_3	50
Cu	CuO	50
Pb	PbO	50
Ag	Ag_2O_2	50
V	V_2O_5	50
Mg	MgO	50
Sc	Sc_2O_3	50
Ni	NiO	50
Al	Al_2O_3	50
Mn	MnO_2	61*
Be	BeO	50
Zn	ZnO	50
Sn	SnO_2	50
	Total	811

*equivalent to 50 mg. of MnO

This material was mixed in an agate mortar for 45 minutes. A mixture containing 1 per cent of each oxide was prepared using feldspar as a base. The same was done for quartz and mica. (This proved to be

a fortunate decision, for notable differences in spectroscopic sensitivities of the various elements from base to base were later proven to exist.) Such a mixture contained 162.2 mg. of the oxide mixture and 837.8 mg. of base mineral. Successively lower concentrations were obtained by the dilution method of Van Tongeren (1938, p. 27), as follows:

	Equivalent per cent of each oxide
837.8 mg. base plus 162.2 mg. oxide mixture	1.00
68.5 " " " 31.5 " 1% "	0.315
" " " " " " 0.315 "	0.100
" " " " " " 0.100 "	0.0315

and so on, to a dilution of 0.000315%.

STANDARD CONCENTRATION CURVES

Standard mixtures were analyzed exactly as were the unknowns. A series of curves (one for each element in each of the three base minerals) was plotted on semilogarithmic paper, with percentage concentrations as ordinates and corresponding Δ log. E values as abscissae. These curves made it possible to convert Δ log. E figures of unknown analyses to actual percentage figures. Small extrapolations beyond the limits of several of the curves were made in a number of conversions.

The standard curves for a feldspar base were used to convert Δ log. E values of unknown feldspar and groundmass analyses to percentages, the curves for a quartz base to convert the unknown quartz analyses, and the curves for a mica base to convert the unknown mica analyses (Fig. 4). Each curve was corrected for contaminations by elements originally present in the mineral. Percentages given are accurate within about 10 per cent.

POSSIBLE ERRORS

Possible causes of error are as follows:

(1) Contamination. Proper precautions were taken.

(2) Nonuniformity of samples. Solid inclusions were eliminated insofar as possible by extreme care in picking the samples. Microscopic fluid inclusions could have been present, in which case the material would have been uniformly mixed throughout the sample during its preparation.

(3) Impurities in the electrodes, already discussed. Some of the electrodes contained a very little Cu.

(4) Adding the internal standard. All weighings were made on an analytical balance, and mixings were thorough.

(5) Lack of uniformity in films is impossible to overcome. Errors from this cause are about 1 per cent, well within allowable limits.

(6) Heavy background on the films. This was overcome by setting the photometer on the background as clear plate, not theoretically exact, but simple and reasonably accurate.

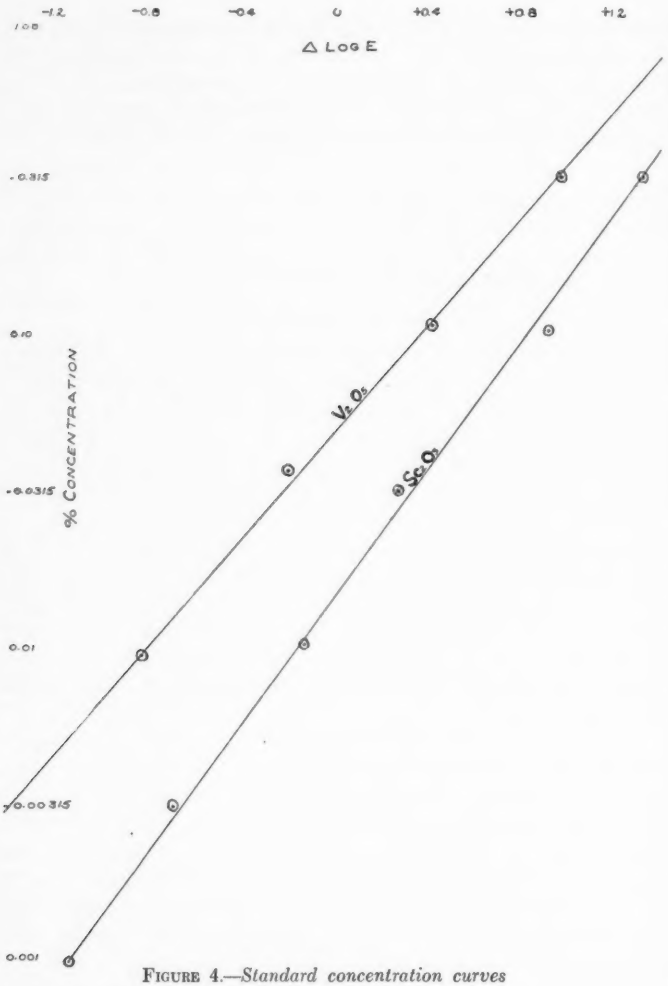

FIGURE 4.—*Standard concentration curves*
(Corrected) for V_2O_5 and Sc_2O_3 in a mica base

(7) Scratches, dust, and spots on the films. These affect density readings. Dust was removed with a soft brush. The films were swabbed several times during and after washing.

(8) Fluctuation in arc conditions. The current and voltage were carefully regulated.

(9) Irregularities in the standard mixtures. The same precautions can be listed here as under (4). Prepared samples were kept in a desiccator to prevent adsorption.

DISTRIBUTION OF MINOR ELEMENTS IN SILICATES

STRUCTURES OF THE SILICATES

For detailed crystal chemistry of the silicates the reader is referred to Stillwell (1938), Wyckoff (1931), Bragg and Bragg (1933), Evans (1939), Bragg (1937), and Hassel (1935). Berman (1937) classified the silicates on a structural basis. A very important contribution was that by Buerger (1934).

The fundamental unit of all silicate structures is the SiO_4 group, the small Si atom at the center of a tetrahedron of larger oxygen atoms. All silicate frameworks are built of these tetrahedra bound together in various ways, a fact that makes most silicate structures more rational than one might judge from the frequently complicated formulae. Thus SiO_4 groups may exist separately, bound together by ions of bivalent positive Mg, for example, which forms the orthosilicate forsterite. Other orthosilicates are similar. If sufficient oxygen is not present to allow formation of independent SiO_4 groups, oxygen atoms may be shared between tetrahedra. Thus each tetrahedron might share one oxygen atom with another tetrahedron, forming the pyrosilicate framework. Other representative frameworks are formed when two, three, or all four oxygen atoms are shared, forming, respectively, the metasilicate ring or chain, the sheet framework of the micas, and the three-dimensional space net of silica. Other modes of sharing are possible.

Trivalent Al can substitute for tetravalent Si in silicates, within certain limits. The Al ion forms both AlO_4 and AlO_6 groups. Substitution of Al for Si often complicates the similar silicate structures, because the lower charge of the Al ion requires one more positive charge to enter the structure in order to balance the negative excess. The degree of substitution of Al for Si often determines the number of additional cations in a structure.

FORM OF THE MINOR ELEMENTS IN THE SILICATES ANALYZED

Solid and fluid inclusions as a source of minor elements have already been mentioned. The minor elements found in the several minerals are probably in solid solution. Interstitial, omission, and substitutional solid solutions can be distinguished. The interstitial type, "stuffing" extra atoms into a crystal lattice, is possible only where sizable openings are present in the various structural units and is therefore limited mainly to elements of small ionic radii. Substitution of one element for another,

very important in the silicates, is the most probable form of solid solution in the minerals analyzed.

The distribution of elements within minerals crystallizing from a magma depends upon (1) the composition of the original liquid, (2) the temperatures and pressures at which the minerals form, and (3) the rate of cooling. Selective distribution of minor elements in solid solution in particular minerals follows certain geochemical principles. The substituting ion should have chemical, physical, and geometrical properties (within a certain range) similar to those of the displaced ion. Ionic size is most important in this respect, although ionic type is also important. Goldschmidt (1937) emphasized the importance of ionic and atomic radii, valence, and ionic type in the formation of crystals, which he likens to sieving mechanisms which grade the elements according to size. Goldschmidt believes that in the appropriate system the larger of two ions is concentrated in the rest liquid and therefore appears in minerals of late formation in that system, or that a crystal may take up ions of similar radius but different charge (of the same sign). In such cases the ion of larger electrostatic charge is given preference over an ion of similar size but lower charge. The plagioclase system is cited as an example, where bivalent Ca is preferentially taken up in early crystals and univalent Na in later crystals. Na and Ca have almost identical ionic radii (Table 5).

The only cation in quartz, Si, is very small, and none of the larger ions are to be expected in substitutional solid solution. Little or no interstitial solid solution would be expected because of the compact quartz structure. Feldspars accept other cations as substitutes for K, Na, Ca, Si, and Al. A striking example is the relative abundance of Ba in microcline, probably as a substitute for K, which has a closely similar ionic radius. Feldspar structures are fairly compact, so little interstitial solid solution, except by very small ions, is to be expected. Foreign elements may substitute in muscovite for K, Al, and Si, as well as the ever-present Fe, Mg, Ca, and Na. The large ions act as a loose "binder" between double sheets of SiO_4 groups bound together by Al atoms, and there is fairly good opportunity between these double sheets for interstitial solid solution. Biotite permits substitution for K, Mg, Fe, Al, and Si. As expected, many more minor elements were found in the micas than in the other minerals analyzed.

Bowen (1928) has given details of crystal fractionation in magmas. If there is a concentration of major elements in a crystallizing melt, according to ionic size or to a series of crystal structure types based on successively less available oxygen, the minor elements should exhibit similar behavior. Thus, if certain minor elements were concentrated in

early minerals by substitution for ions of the proper size, these same elements should decrease in quantity in later minerals, even though favorable conditions then exist for their acceptance in the formative crystals. The same might be said concerning minerals in an igneous rock and the same minerals in associated pegmatites. Thus, if certain minor elements segregated in particular minerals during early crystallization, the same minerals formed at the pegmatite stage would probably contain less of these elements, which would be less abundant in the pegmatitic liquid. A much more plausible reason for this condition is that at higher temperatures when the structures are in a state of "dynamic disorder" (Buerger, 1934; also personal communication) crystals are more tolerant to substitutional solid solution than at lower temperatures, when the structures have become more ordered. Since pegmatites form at lower temperatures, the ordered mineral structures are not so tolerant to foreign elements as were the same, but disordered, structures at the higher parent-rock temperatures. This predictable relation of parent rock to pegmatite was strikingly borne out by the analyses.

An excellent foundation for this work can be gained from the literature. Of special significance here are the works of Goldschmidt and his co-workers (1929 to present), Noll (1934), Von Engelhardt (1936), Schröder (1932), Tröger (1935), Van Tongeren (1938), Harcourt (1934), Knopf (1936), Kennard (1935), and Glass (1935).

RESULTS OF QUALITATIVE ANALYSES OF ROCK SAMPLES

GENERAL

Tables 2, 3, and 4 list results of qualitative rock analyses. Figure 1 gives the locations of the samples. Analyses of two Massachusetts rocks (Table 4) are included for comparison with the Jamestown results. The symbols listed in Table 2 have already been explained. These symbols should be used only in comparisons involving a single element.

The major elements Si, Al, Fe, Mg, Ca, Na, and K need no discussion. They are important in that they control the distribution of certain of the minor elements. Table 5 lists the ionic radii of elements most vital to the discussion. Descriptive terms will refer only to spectroscopic quantities. Thus "large" means large only from the spectroscopic viewpoint.

DISTRIBUTION OF THE MINOR ELEMENTS IN THE ROCKS

Ba and Sr are present in all, with Sr distinctly less in the Massachusetts rocks than in most from Colorado. All the rocks were acid, so no striking differences in content of Ba and Sr should be expected.

Sc is generally present in the Colorado rocks (highest in the grano-

TABLE 2.—*Qualitative analyses of pre-Cambrian rocks*

Granite	Sample number	Mg	Sr	Ba	Sc	Y	La	Ce	Nd	Ti	Zr	V	Cr	Mn	Fe	Co	Ni	Ag	Zn	Pb
	58	O	VL	VL+	S+	A	A	A	?	L−	T−	L	L	L	L+	VT	S−	A	A	A
	82	O	VL	VL+	S+	A	A	A	?	L−	T−	L	L+	L	L+	T	S+	A	A	A
Boulder	81	VL	VL−	VL−	T	A	A	A	O	L	A	L−	M+	L−	L	VT	T	A?	A	A
Creek	54	VL	L+	VL	T	A	A	A	O	L	T−	L	M	M	L	VT	VT	A	A	VT−
	74	VL	VL−	VL	T+	A	A	A	O	L	T−	L−	M	M	L	VT	VT	A	A	A
	79	VL	L+	VL	VT	A	VT	VT	O	L	T−	L−	M	M+	L	VT	VT	A	A	A
	85	VL	L+	VL	T+	VT+	A	A?	O	L	VT	L−	M	L−	L	VT	VT	VT	A	A
	106	O	L	VL	T	A	VT	T	T	M+	T	M+	S+	M	O	A	A	A	A?	A
Overland	110	VL	M	M	A	A	A	A	T	L	A	L	T+	M	L	A	A	T−	VT?	A
Mountain	113	VL	M+	L	S−	A	S	VT+	T+	L	VT	M−	S−	M	L	A	A	VT	A	A
	108	VL	L	VL	S−	A	A	VT	T?	L	A	M	T+	M	L	VT−	A	A	A	VT
	104	O	O	VL	VT	A	T	T	T+	L	T	L	M	L	O	VT	VT?	A	A	T+
	117	O	O	VL	VT	A	A	VT	A?	L	T+	M	M	L	O	VT−	VT?	A	A	T
Silver	120	O	O	VL	A	A	VT	T	A	L	T+	M	S+	L	O	A	A	A	A	T
Plume	98	O	O	VL	VT	VT	S	T	T	L	M−	M	S+	L	O	A	A	A?	A	T
	122	VL	VL	VL+	VT	A	VT+	VT+	VT	L	T+	M+	M−	M	L	A	A	A	A	VT−
	100	VL	VL	VL+	VT	VT+	T	VT+	VT	L	VT	M	S+	M	L	A	A	A	A	VT
	124	VL	L	VL	A	A	VT	VT+	VT	M+	VT−	M	S+	M−	L	A	A	VT	A	VT−
	102	VL	L	VL−	VT	A	A	A	VT−	L	VT	M	M−	M−	L	A	A	A	A	A

Not detected in any samples: Sb, As, Bi, Cd, Be, Ge, Au, Os, Pd, Pt, and Sn.

Legend

Symbol	Spectroscopic Quantity present (no definite percentage)			
O	not determined	S	small	(plus and minus
A	not detected	M	medium	signs show variations
VT	very small trace	L	large	within the ranges given)
T	trace	VL	very large	

TABLE 3.—*Qualitative analyses of Tertiary Colorado rocks*

Rock	Sample number	Mg	Sr	Ba	Sc	Y	La	Ce	Nd	Ti	Zr	V	Cr	Mn	Fe	Co	Ni	Ag	Zn	Pb
	87	O	VL	VL	S−	A	A	VT	A	L	A	L	M−	L	L+	A	A	A	A	A
	26	O	VL	VL	S−	A	A	A	A	L	T	L	M	L	L+	VT	A	A	A	A
	17	O	L+	L+	VT	A	A	A	A	M+	A	M	S	L	O	VT	A	A	A	A
Grano-	40	O	VL	VL	S+	A	A	A	T?	L	A	L−	S	M+	L	T	VT	A	T−	A
diorite	46	O	VL	VL	S−	A	T	VT−	T?	L	A	M+	S	M+	L	VT	VT	VT	A	A
	15	O	VL	VL	S	A	VT	A	T	L	S	L	M−	L+	O	VT+	VT?	A	A	A
	50	O	VL	VL	S+	A	VT	A	T	L	VT	L	M−	L+	O	VT+	VT?	A	A	A
	29	O	VL	VL	S+	A	VT	A	T	L	VT	L	M−	L+	O	VT	VT?	A	A	A
	33	O	VL	VL	S	A	A?	A	T	L	A	L	M−	L+	O	VT	A	A	A?	A
	7	O	VL	VL	A	A	A	T	A	M	A	L−	S	S+	L+	A	A	A	A	A
	1	O	VL	VL	A	A	A	T	A	M	A	L−	S	S+	L	A	S	A	A	A
Granite-	5	O	L+	L+	A	A	A	A	A	S+	A	S	VT	L	O	A	A	A	A	A
quarts	11	L	VL	VL	VT	A	A	A?	O	M+	A?	L−	T+	M	L	A	A	A	A	VT−
monzonite	3	L	L−	VL	A	A	A	A?	O	M+	A	L−	S+	S	L	A	A	A	A	A
	116	L+	VL	VL	VT	A	A	A?	O	L−	VT	M+	S+	T+	L	A	A	A	A	A
	9	L+	VL	VL	A	VT+	A	A?	O	M	A	M+	T	S+	L	A	A	VT	A	A?

Symbols as in Table 2.
Not detected in any samples: Sb, As, Bi, Cd, Be, Ge, Au, Os, Pd, Pt, and Sn.

TABLE 4.—*Qualitative analyses of Massachusetts rocks (not Tertiary)*

Rock	Mg	Sr	Ba	Sc	Y	La	Ce	Nd	Ti	Zr	V	Cr	Mn	Fe	Co	Ni	Ag	Zn	Pb
Fine Quincy granite	S+	T+	L	A	VT	A	A	O	M+	S	T?	VT−	L−	L	A	A	A	VT	A
Coarse Quincy	S+	M	L	A	VT	A	A	O	M+	S+	T?	VT−	L	L	A	A	VT	T	VT+
Coarse Dedham gnd.	L+	L	VL	A	A	A	A	O	M	VT	M−	VT	L−	L	A	A	A	A	A
Typical Dedham	VL	L	VL	T	T	A	A	O	L−	VT	M	VT	L	L	A	A	A	A	A

Symbols as in Table 2
Not detected in any samples: Sb, As, Bi, Cd, Be, Ge, Au, Os, Pd, Pt, and Sn.

diorite) but not in those from Massachusetts. It is highest where Mg and ferrous Fe are abundant, because of their similar ionic radii (Table 5).

Y is normally absent in the pre-Cambrian and sporadic in the Tertiary Colorado rocks but was detected in almost all Massachusetts samples.

TABLE 5.—*Ionic radii of elements in the silicates*

(*See* Stillwell, 1938, Appendix)

Element	Group of periodic system	Atomic number	Atomic weight	Valence	Ionic radius
Na	I	11	22.99	1	0.98 Å
K	I	19	39.10	1	1.33
Mg	II	12	24.32	2	0.78
Ca	II	20	40.08	2	1.06
Sr	II	38	87.63	2	1.27
Ba	II	56	137.36	2	1.43
Al	III	13	26.97	3	0.57
Sc	III	21	45.10	3	0.83
Y	III	39	88.92	3	1.06
La	III	57	138.90	3	1.22
Ce	III	58	140.13	3	1.18
Nd	III	60	144.27	3	1.17
Si	IV	14	28.06	4	0.39
Ti	IV	22	47.90	4	0.64
Zr	IV	40	91.22	4	0.87
V	V–B	23	50.95	3	0.75
V	V–B	23	50.95	4	0.61
V	V–B	23	50.95	5	
Cr	VI–B	24	52.01	3	0.65
Mn	VII–B	25	54.93	2	0.91
Fe	Transition	26	55.84	2	0.83
Fe	Transition	26	55.84	3	0.67
Co	Transition	27	58.94	2	0.82
Ni	Transition	28	58.69	2	0.78
Cu	I–B	29	63.57	1	0.96
Zn	II–B	30	65.38	2	0.83
Pb	IV–B	82	207.22	2	1.32

La is almost universal in the Silver Plume samples and common in the granodiorite. Ce occurs frequently in the Overland Mountain and (especially) Silver Plume granites. It is notable that several cerite deposits, associated with the Silver Plume, are present in the Jamestown district (Goddard and Glass, 1940). Ce is also present in the Tertiary granite. The rare earths are thought to occur as substitutes for Ca and Na, because of similarity in ionic radii. Tröger (1935) gave the average rare-earth contents of several minerals, of which sphene, zircon, and apatite were highest.

Ti, detected in all rocks, occurs in accessories (ilmenite, sphene, and rutile) and as a substitute in rocks high in Mg ferrous Fe. Most samples contain Zr, the Quincy (Mass.) granite most of all, and the Tertiary stocks least. The Zr content of the rocks is probably a function of the amount of zircon present.

TABLE 6.—*Qualitative scheme of "index elements"*

Rock	Sc	La	Ce	Nd	Ti	Zr	V	Cr	Co	Ni
Pre-Cambrian Colorado rocks										
Silver Plume*	VT	VT–T	VT–T	VT–T	L	T	M	S–M	A–VT	A
Overland Mtn.	T	Frequent	Frequent	T	L	A–T	M	S	A	A
Boulder Creek	T–S	A	A	O	L	T	L	M–L	VT	VT–T
Tertiary Colorado rocks										
Granite-quartz monzonite	A–VT	A	T?	A	M	A	M–L	S	A	A
Granodiorite	S	Sporadic	A	Sporadic	L	VT	L	S–M	VT	A–VT
Massachusetts rocks										
Quincy**	A	A	A	O	M	S	T?	VT	A	A
Dedham***	A	A	A	O	M–L	VT	M	VT	A	A

Symbols as in Table 2.

* Pb in all samples (VT).
** Also contains Zn.
*** Also contains Y (T).

V, found in all samples, is most abundant in those rich in ferric Fe. It is highest in the Boulder Creek granite (pre-Cambrian) and the granodiorite (Tertiary). There are a number of pyritic veins near Jamestown that bear the V mica roscoelite, believed to have formed immediately following the implacement of the granodiorite. The Massachusetts rocks are poorer in V than are those from Colorado.

Cr, also found in all samples, is highest in those where ferric Fe is abundant. The Boulder Creek granite is highest in Cr for the pre-Cambrian series, and the granodiorite for the Tertiary rocks. The Massachusetts rocks are poor in Cr. It is notable that V and Cr are distributed similarly; both substitute for ferric Fe because of similarity in ionic radii (Table 5).

Co and Ni are practically restricted to the Boulder Creek granite of the pre-Cambrian group and to the Tertiary granodiorite and are thus highest where biotite and hornblende are abundant, probably occurring as substitutes for Mg and ferrous Fe, which have similar ionic radii. Some Ni may occur in very finely disseminated pyrrhotite and pentlandite (Newhouse, 1936).

Ag was found in several samples. Zn occurs in both Quincy granite samples and in one of granodiorite. Pb is present in the Silver Plume granite but is otherwise rare. Since Pb is present in the youngest pre-Cambrian rock and absent in the others, we can scarcely say that the Pb is due to radioactive disintegration, for in such a case Pb would be least abundant in the younger rocks.

SIGNIFICANCE OF RESULTS

Each rock is characterized by certain minor elements. If the rocks are arranged according to age and province, as in Table 6, it is possible to set up a scheme whereby each rock can be distinguished from all others. The elements that show significant differences from rock to rock are termed "index elements." By their use one should be able to take a powdered sample of any of the rocks concerned, analyze it spectrographically, and identify it. Differences between the analyses of rocks from the two provinces are especially notable.

Lovering and Goddard (1939) classify the Overland Mountain granite body as Boulder Creek. No doubt these rocks look very similar, are of the same igneous sequence, and would be so indicated on a map, but certain differences in content of minor elements (Sr, Zr, rare earths, V, Cr, Co, and Ni) give some basis for a local division.

QUALITATIVE WORK ON MINERAL SAMPLES

GENERAL

The qualitative work on minerals was done differently from that on rocks, so the two sets of results are in no way comparable. Tables 7, 8, 9, and 10 give the results for mineral analyses. The minerals analyzed include potash and plagioclase feldspar, muscovite and biotite, quartz, sphene, hornblende, magnetic concentrates, and groundmass. The results for magnetic concentrates are not so desirable as one might wish, because the samples tended to hop from the electrodes during excitation.

DISTRIBUTION OF MINOR ELEMENTS ACCORDING TO MINERAL SPECIES

Sr is highest in potash feldspars and hornblendes, high in plagioclases, groundmasses (samples of groundmasses of the porphyritic rocks), and sphenes, and medium in micas. Ba is highest in potash feldspars and groundmasses, abundant in the micas and plagioclases, and medium in sphene and hornblende. Since Ba is highest in high-K minerals, the Sr:Ba ratio is probably higher in minerals richer in Ca, such as plagioclase and sphene, than in high-K minerals. This principle was formulated by Noll (1934). The explanation lies in the ionic radii of the elements concerned (Table 5). Sr is more readily interchangeable with Ca than

is Ba. Ba is much more readily interchangeable with K, a fact that aids in explaining the high Ba content of K minerals. Sr also enters K minerals, because the Sr ion is smaller than that of K. The sphenes are rich in Al, probably because Al and Ti have similar ionic radii (Table 5). Al is also present in quartz. The Al in magnetic concentrates may be due to intergrowths of spinel in the opaque oxides. Sc is high in hornblendes, medium in micas, and low in sphenes and groundmasses. As a minor element Sc is most abundant where Mg and ferrous Fe are important major elements, as in hornblende or mica, because of similarities in the ionic radii.

Y occurs only in the sphenes. La is present in sphenes, groundmasses, and in the Silver Plume biotite. Ce, like La, occurs almost exclusively in the sphenes and groundmasses. Nd is present in all sphene samples and in the hornblendes. The most important point here is the marked preference of the rare earths for the mineral sphene (and to a lesser extent the groundmasses), probably as substitutes for Ca (Tröger, 1935). The granodiorite magma was apparently rich in the rare-earth elements, which were presumably taken up in the early-formed, high-Ca minerals.

Ti is high in the biotites, groundmasses, muscovites, and hornblendes. Zr, never abundant, occurs only in hornblendes, biotites, and several muscovites, probably as a substitute for Mg and ferrous Fe. V is highest in the sphenes and hornblendes, less abundant in biotites and magnetic concentrates, and least in groundmasses and muscovites. It probably substitutes for Ti in sphene and hornblende and for ferric Fe in micas and magnetic concentrates. Cr is distributed exactly as is V, and for the same reasons just given for V. Mn occurs in practically all samples except quartz, most abundantly in sphene, the micas, hornblende, and the groundmasses. Mn probably substitutes for ferrous Fe. Ni and Co are present in several samples of high-Fe minerals. Zn, with an ionic radius identical to that of ferrous Fe, was detected in most magnetic concentrates. Table 11 lists the minerals with their characteristic minor elements.

DIFFERENCES IN CONTENT OF MINOR ELEMENTS IN SINGLE MINERALS

Potash feldspar (Table 7).—Tertiary potash feldspars are poorer in Ba than are the pre-Cambrian. Mn is highest in the Silver Plume and Tertiary samples. The parent rock-pegmatite relations are summarized as follows:

Boulder Creek granite			Silver Plume granite		
Parent rock		Pegmatite	Parent rock		Pegmatite
More	Mg	Less	More	Mg	Less
More	Sr	Less	More	Sr	Less
More	Ba	Less	More	Ba	Less
More?	Ti	Less?	More	Ti	Less
More?	Mn	Less?	More	Mn	Less

TABLE 7.—*Qualitative analyses of feldspars*

Rock	Sample number	Na	K	Mg	Sr	Ba	Nd	Ti	Cr	Mn	Fe
				Potash Feldspars							
(Pre-Cambrian)	P.R. 79	S	O	S	L+	VL+	A	VT	A	A	A
	85	M	O	S+	VL	VL+	A	VT	A	A	A
Boulder Creek granite	Peg. 59	M	O	S−	VL	VL+	A	VT	A	A	A
	76	M	O	VT	M	M+	A	A	A	T	A
Overland Mountain granite	P.R. 110	M+	O	M	VL	L+	A	VT	A	M	Present
	108	M	O	S	VL	VL+	A	S	A	VT	A
	104	M+	O	M	L+	VL+	A	T	A	VT	Present
	P.R. 122	M	O	M	L+	VL+	A	S	A	S	Present
	120	M	O	M+	L+	VL+	A	S	A	T	Present
Silver Plume granite	98	M	O	M+	L+	VL+	A	S	A	T	Present
	Peg. 88	M+	O	S	S	M	A	A	A	A	A
	95	M	O	S	M	M	A	VT	A	A	A
(Tertiary)											
Tertiary granodiorite	50	M	O	M	VL	VL	A	M+	A	S	Present
Granite-quartz monzonite	11	M+	O	S	VL	VL	A	T	A	T	Present
				P.R.—Parent rock; Peg.—Pegmatite.							
				Plagioclases							
Boulder Creek granite	79	O	S	M	L	L	A	S+	A	T	Present
	85	O	M	M	L	VL	A	S+	A	S+	Present
Tertiary granodiorite	33	O	S	M+	VL	VL	A	S+	A	T	Present

Symbols as in Table 2

Not detected in any samples: Sb, As, Be, Bi, B, Cd, Ce, Cs, Co, Cu, Ge, Au, Hf, In, Ir, La, Pb, Mo, Ni, Os, Pd, Pt, Sc, Ag, Sn, W, V, Y, Zn, and Zr.

Thus the pegmatite feldspars bear less of practically every minor element present than those from the parent rocks, a fact particularly striking for the Silver Plume granite.

Plagioclase, groundmass, and quartz (Tables 7, 9, and 10).—Data for plagioclase and groundmass samples are too meager to permit conclusions.

Mg, present in parent-rock quartz, is absent in that from pegmatites. For the Silver Plume granite Ca, Ba, Al, Ti, and Fe are much less abundant in pegmatite quartz than in parent-rock quartz. The same is true for Fe and Mn in the case of the Boulder Creek granite. The reverse seems to be true, however, for Ca, Ba, and Al. (It must be borne in mind that the data here are very meager.)

Micas (Table 8).—Too few data are available for the Overland Mountain granite muscovite. The parent rock-pegmatite relations for the Boulder Creek and Silver Plume muscovites are as follows:

Boulder Greek granite			Silver Plume granite		
Parent rock		Pegmatite	Parent rock		Pegmatite
Same	Na	Same	More	Na	Less
Same	Mg	Same	More	Mg	Less
More	Sr	Less	More	Sr	Less
More	Ba	Less	More	Ba	Less
More	Sc	Less	Same	Sc	Same
More	Nd	Less	More	V	Less
More	V	Less	Same	Ti	Same
Same	Ti	Same	Some	Cr	None
Some	Zr	None	Less?	Mn	More?
More	Cr	Less			
Less?	Mn	More?			

The Boulder Creek muscovites contain more Sr, Ba, Sc, Ti, Zr, V, Cr, and Mn than those from the Silver Plume.

The Boulder Creek biotites bear more Sc, V, Cr, and Mn than those from any other rock. La occurs only in the Silver Plume biotite, where Nd and Zr are also most abundant. Ni is confined to the Boulder Creek biotites.

Hornblende, sphene, and magnetic concentrates (Tables 9 and 10).—The hornblendes are notable for their content of Zr and high concentration of Sc. They also bear high Mn, V, and Cr. The sphenes are notable for their high rare-earth content, as well as high Al, Fe, Mn, Cr, and V.

The Tertiary magnetic concentrates contain more Mg, Ti, V, Mn, and Zn than those from pre-Cambrian rocks. The Boulder Creek samples bear more Mg, Ca, Ba, Al, Ti, V, Cr, and Mn than that from the Overland Mountain granite.

TABLE 8.—*Qualitative analyses of micas*

Rock	Sample number	Na	K	Mg	Sr	Ba	Sc	La	Ce	Nd	Ti	Zr	V	Cr	Mn	Co	Ni	Zn
							Muscovites											
Boulder Creek granite	P.R. 85	S	VL	VL	L	VL+	M+	A	A	VT	L	VT	L	M+	L	A	A	A
	Peg. 76	S	VL	VL	S	VL	M−	A	A	A	L	A	VT	A	L+	A	A	A
Overland Mtn. granite	108	T	VL	VL	M	VL+	L	A	A	A	L	S	M	S	L+	A	A	A
Silver Plume granite	P.R. 122	VT	M	VL	VT	VL	T	A	A	A	M+	A	S	S	S	A	A	A
	98	S	VL	VL+	M	VL+	M−	A	A	A	M+	A	M−	S	M	A	A	A
	Peg. 95	VT	VL	VL	T	L+	M−	A	A	A	M+	A	S	A	M+	A	A	A

P.R.—Parent rock; Peg.—Pegmatite

							Biotites											
Boulder Creek granite	79	VT	VL	VL+	M+	VL	M	A	A	VT	L−	A	M+	L	VL	A	VT	A
	85	A	M	VL+	S	VL	M	A	A	VT	L	S	M+	L−	L	A	VT	A
Overland Mtn. granite	108	A	VL	VL+	M	VL	M	A	A	VT	L	VT	M	S	VL	A	A	VT
Silver Plume granite	104	A	L	VL+	A	VL	S	VT	A	VT	L	S	M	M−	M	VT	A	A
	122	VT	VL	VL+	M	VL	S	M−	T	S	L	S+	M+	M	VL	A	VT	A
	120	S	VT	VL+	A	L	VT	A	A	A	L	VT	VT	T	M	A	A	A
	98	VT	L	L	M	VL	A	T	A	T	L	S	S	T	M	A	A	A

Symbols as in Table 2

Not detected in any samples: Sb, As, Be, Bi, B, Cd, Cs, Ge, Au, In, Ir, Mo, Os, Pd, Pt, Ag, Sn, W, and y.

TABLE 9.—*Qualitative analyses of quartz and magnetic concentrates*

Rock	Sample number	Na	K	Mg	Ca	Sr	Ba	Al	Ti	V	Cr	Mn	Fe	Zn
						Quartz								
	P.R. 85	A	A	M	A	A	A	S	A	A	A	M	T	A
Boulder Creek granite	59	A	A	A	M	A	A	L	A	A	A	A	A	A
	Peg. 76	A	A	A	VT	A	VT	S	A	A	A	A	A	A
Overland Mtn. granite	110	A	A	T	M+	A	VT	L−	T	A	A	A	T	A
	120	A	A	S+	L	S	L	VL+	T	A	A	T	M−	A
	P.R. 98	A	A	S+	L	A	L	VL	T	A	A	A	S+	A
Silver Plume granite	88	A	A	A	VT	A	VT	S	A	A	A	A	A	A
	Peg. 95	A	A	A	M	A	VT	M+	A	A	A	A	A	A

P.R.—Parent rock; Peg.—Pegmatite

						Magnetic Concentrates								
Boulder Creek granite	82	A	A	T	L+	A	M	L	S	S	S	T	O	A
	58	A	A	T	L+	A	M	L	S	M	M+	T	O	A
	81	A	A	A	M	A	M	M	VT	A	VT	A	O	A
Overland Mtn. granite	113	A	S+	A	M	A	VT	L−	VT	A	T	VT	O	A
Granite-quartz monzonite	11	VT	S	T	M+	A	VT	L−	T+	M	T	M	O	VT
	9	S	M	T	M+	A	VT	L−	T+	M	T	S	O	A
Tertiary granodiorite	40	A	A	T	L	A	T	M	S	S	A	M	O	VT
	33	A	A	T	L	A	T	M−	T+	S	A	S	O	VT
	87	A	A	S	L	A	T	M+	T+	M	S+	M	O	A

Symbols as in Table 2

Not detected in any samples: Sb, As, Be, Bi, B, Cd, Cs, Co, Ge, Au, In, Ir, La, Pb, Mo, Ni, Os, Pd, Pt, Ag, Sn, W, Y, and Zr.

TABLE 10.—*Qualitative analyses of sphene, hornblende, and groundmass*

Rock	Sample number	Na	K	Mg	Sr	Ba	Al	Sc	Y	La	Ce	Nd	Ti	Zr	V	Cr	Mn	Fe	Co	Ni
									Sphene											
	40	T	M	M	L	M	VL+	S+	S+	L−	M	M	VL	A	L	M	L	L+	A	A
Tertiary	50	T	M	M	L	M	VL+	S+	M−	L−	M	M	VL	A	L	M	L	L+	A	A
Granodiorite	33	VT	A	M	M	S	VL	T+	T	M	S+	S	VL	A	L	T+	L	L	A	A
	17	VT	S	M	M	M	VL	T+	T+	M	S+	S+	VL	A	L	S	L	L	A	A
									Hornblende											
	40	S	M	VL+	VL	M	O	M+	A	A	A	VT?	M+	VT	M	S+	M+	O	A	A
Tertiary	50	S	M	VL+	VL	M	O	L	A	A	A	VT?	L−	S	L	L−	M+	O	A	A
Granodiorite	33	S	M	VL+	VL	M	O	L	A	A	A	T	M+	A	M	T	M+	O	A	A
	87	S	M	VL+	VL	L	O	L	A	A	A	T	L	VT	L	M	M+	O	S	VT
	17	T+	S	VL+	L	L	O	L	A	VT−	A	T	L+	VT	L	M	M+	O	S	A
									Groundmass											
Granite-quartz monzonite	11	M+	VL	M	VL	VL+	O	VT	A	T	VT	A	L	A	M	VT	M	O	A	A

Symbols as in Table 2

Not detected in any samples: Sb, As, Be, Bi, B, Cd, Cs, Ge, Au, In, Ir, Pb, Os, Pd, Pt, Ag, Sn, and W.

SUMMARY OF QUALITATIVE WORK

ROCKS

Qualitative analyses were made the basis for a scheme whereby any rock concerned could be identified upon analysis of its powder (Table 6).

TABLE 11.—*Minerals, with characteristic minor elements*

Potash feldspar	Plagioclase	Quartz	Muscovite	Biotite
Mg — S-M	Mg — M	Mg — Spor.*	Na — T-S	Na — VT
Sr — VL	Sr — L	Ca — M	Mg — VL+	Sr — S
Ba — VL+	Ba — L	Ba — Spor.	Sr — M	Ba — VL
Ti — VT-S	Ti — S	Al — M	Ba — VL	Sc — T
Mn — T	Mn — T	Ti — Spor.	Sc — M	Nd — T
		Fe — Spor.	Ti — L	Ti — L
			V — M	Zr — S
			Cr — S	V — M
			Mn — L	Cr — M
				Mn — L

Hornblende	Sphene	Magnetic concentrate	Groundmass
Sr — VL	Na — T	Na — Spor.	Sr — VL
Ba — M	Mg — S	K — Spor.	Ba — VL+
Sc — L	Sr — L	Mg — T	Sc — VT
Nd — T	Ba — M	Ca — L	La — T
Ti — L	Al — VL	Ba — S	Ce — VT
Zr — VT	Sc — S	Al — L	Ti — L
V — L	Y — S	Ti — S	V — S
Cr — M	La — M	V — M	Mn — L
Mn — L	Ce — M	Cr — S	
Co — Spor.	Nd — M	Mn — S	
	V — L	Zn — VT	
	Cr — M		
	Mn — L		
	Fe — L		

Symbols as in Table 2

* Sporadic.

Ba and Sr are abundant in all rocks, because of the tendency of Ba to substitute for K, and Sr for Ca and K. Sc substitutes for Mg and ferrous Fe. This is in accord with the work of Goldschmidt and Peters (1931b). Ti occurs in all rock samples. Cr and V, found in all samples, probably substitute for ferric Fe in the dark minerals. Ni and Co are distributed in much the same manner as are V and Cr. The following elements were not detected: As, Sb, Bi, Cd, Be, Ge, Au, Os, Pd, and Pt.

The Colorado rocks differed from the Massachusetts rocks in a number of important respects, thus emphasizing the province relationship.

MINERALS

Sr is high in Ca- or K minerals, and Ba in K minerals. The Sr:Ba ratio is highest in Ca minerals. Al is high in sphene and occurs in all

TABLE 12.—*Quantitative analyses of potash feldspars*

Rock	Sample number	Δ log. E Mg	Δ log. E Ca	% SrO	% BaO	% TiO_2	% V_2O_3	% Cr_2O_3	% MnO	Δ log. E Fe
					Pre-Cambrian					
	58–3	0.64	0.29	0.065	0.160	0.0140	N.D.	N.D.	0.0048	0.12
	P.R. 82–3	0.85	0.38	0.085	0.100	0.0155	N.D.	0.0003	0.0175	0.41
	81–3	0.58	0.05	0.073	0.018	0.0165	N.D.	N.D.	0.0205	0.21
	54–3	0.31	0.11	0.113	0.260	0.0062	N.D.	N.D.	N.D.	−0.51
Boulder Creek granite	74–3	0.65	0.26	0.100	0.165	0.0100	N.D.	N.D.	0.0100	0.00
	61–1	0.09	−0.25	0.040	0.022	N.D.	N.D.	N.D.	0.0185	−0.42
	Peg. 56–1	−0.35	−0.06	0.040	0.0088	N.D.	N.D.	N.D.	0.0062	−0.62
	83–1	0.21	−0.11	0.052	0.041	N.D.	N.D.	N.D.	0.0026	−0.48
	P.R. 106–3	0.23	0.20	0.090	0.235	0.0320	N.D.	N.D.	0.0034	−0.44
	113–3	0.28	0.10	0.170	0.093	0.0210	N.D.	N.D.	0.0073	−0.23
Overland Mtn. granite	109–1	−0.45	0.17	0.096	0.125	0.0070	N.D.	N.D.	0.0029	−0.68
	Peg. 112–1	−0.03	−0.03	0.160	0.053	N.D.	N.D.	N.D.	N.D.	−0.31
	117–2	0.49	−0.12	0.091	0.066	0.0215	N.D.	N.D.	0.0075	−0.13
	P.R. 124–2	0.48	0.28	0.064	0.130	0.0240	N.D.	N.D.	0.0068	−0.04
	100–2	0.39	0.00	0.105	0.053	0.0115	N.D.	0.0024	0.0075	−0.25
Silver Plume granite	102–2	0.65	0.19	0.065	0.130	0.0210	N.D.	N.D.	0.0040	0.15
	90–1	0.31	−0.16	0.076	0.026	N.D.	N.D.	N.D.	0.0060	−0.50
	Peg. 89–1	0.16	−0.01	0.046	0.048	N.D.	N.D.	N.D.	0.0040	−0.82

P.R.—Parent rock; Peg.—Pegmatite. N.D.—not detected

					Tertiary					
	87–3	0.87	0.56	0.0970	0.160	0.0410	N.D.	N.D.	0.0044	0.47
	40–3	0.79	0.49	0.1165	0.105	0.0340	N.D.	0.0002	0.0056	0.34
Granodiorite	33–3	0.49	0.02	0.220	0.058	0.0170	0.0033	0.00023	0.0075	0.01
	17–3	0.59	−0.01	0.123	0.037	0.0148	N.D.	0.00021	0.0096	−0.01
	9–2	0.27	0.21	0.160	0.140	0.0090	N.D.	N.D.	0.0031	0.25
Granite-quartz monzonite	3–3	0.37	−0.09	0.295	0.120	0.0260	0.0040	0.00044	0.0135	0.08
	116–3	0.74	−0.02	0.245	0.130	0.0500	0.0025	N.D.	0.0025	−0.03

N.D.—not detected

quartzes. Sc substitutes for Mg and ferrous Fe in the dark minerals. Sphene bears a notable quantity of rare earths. Hornblende contains Nd. Ti is highest in the micas, groundmasses, and hornblende. Cr and V are distributed in exactly the same manner, mainly in the dark minerals as substitutes for ferric Fe. Mn substitutes for ferrous Fe in the dark minerals, as do Co and Ni. Zn was detected only in the Tertiary magnetic concentrates. The following elements were not detected: Sb, As, Be, Bi, B, Cd, Ge, Au, In, Ir, Mo, Os, Pd, and Pt.

Pegmatite minerals usually contain less of almost every minor element than the same minerals from the parent rock. This can be explained on the basis of the temperature-crystal structure-composition relations of crystals forming in an igneous melt. This must not be construed to mean that the minor elements are necessarily less abundant in the pegmatitic stage of magmatic history, for it is well known that some actually concentrate then. However, the bulk of the minor elements may not be absorbed in the common pegmatite minerals, because of lowered temperature and pressure in the (essentially) open system. They may escape instead, perhaps to form small quantities of minerals (in the pegmatite or a vein) in which the minor constituents of the pegmatite system as a whole might appear as essential elements, as in the Li minerals. On the other hand, the magma system is essentially closed, and the rare elements are forced to enter the formative crystals. All the rocks can be distinguished on the basis of mineral analyses.

QUANTITATIVE WORK ON MINERALS

GENERAL

Quantitative spectrographic analyses of mineral samples supported the conclusions drawn on the basis of qualitative work. The quantitative results are given in Tables 12, 13, 14, 15, and 16 as percentages of oxides or, where actual percentages were unobtainable, as Δ log. E figures. Negative Δ log. E figures indicate less of an element than do positive figures. They are definitely quantitative but should be used only in comparisons involving a single element in each of the following restricted mineral groups: (1) quartz, (2) feldspars and groundmasses, and (3) micas and hornblendes. The elements are arranged in the tables according to their positions in the periodic classification of elements.

DISTRIBUTION OF MINOR ELEMENTS ACCORDING TO MINERAL SPECIES

General statement.—Mg is a minor element in feldspars and some quartzes, and Ca is present in almost all minerals. Feldspar and groundmass samples are rich in SrO. The micas bear less. Potash feldspars are

TABLE 13.—*Quantitative analyses of plagioclases and groundmasses*

Rock	Sample number	Δ log. E Mg	% SrO	% BaO	% Sc_2O_3	% TiO_2	% V_2O_3	% Cr_2O_3	% MnO	Δ log. E Fe
					Plagioclase—Pre-Cambrian					
	58–2	0.65	0.045	0.0188	N.D.	0.0033	N.D.	N.D.	0.0064	0.09
	82–2	0.64	0.093	0.069	N.D.	0.0066	N.D.	N.D.	0.0114	0.28
Boulder Creek granite	81–2	0.59	0.070	0.009	N.D.	0.0170	N.D.	N.D.	0.0120	0.02
	54–2	0.50	0.029	0.0044	N.D.	0.0053	N.D.	N.D.	0.0370	−0.31
	74–2	0.58	0.093	0.069	N.D.	0.0078	N.D.	N.D.	0.0170	−0.01
Overland Mtn. granite	106–2	0.48	0.041	0.115	N.D.	0.0070	N.D.	0.0001	0.0028	−0.62
	113–2	0.54	0.050	0.0055	N.D.	0.0090	N.D.	N.D.	0.0052	−0.11
					Plagioclase—Tertiary					
	87–2	0.57	0.073	0.0071	N.D.	0.0180	0.001	N.D.	0.0050	0.22
Granodiorite	40–2	0.67	0.231	0.0205	N.D.	0.0380	N.D.	N.D.	0.0190	0.50
	50–2	0.53	0.082	0.0041	N.D.	0.0180	0.001	N.D.	0.0120	0.32
	17–2	0.72	0.093	0.013	N.D.	0.0180	N.D.	N.D.	0.0160	0.27
					Groundmass—Tertiary					
Granite-quartz monzonite	9–3	0.54	0.105	0.026	0.0022	0.180	0.009	0.0002	0.0148	1.06
	116–4	0.74	0.124	0.087	N.D.	0.370	0.015	0.00022	0.0027	0.49

N.D.—not detected

TABLE 14.—*Quantitative analyses of quartz (from pre-Cambrian granites)*

Granite	Sample number	Δ log. E Mg	% BaO	% Al_2O_3	% TiO_2	Δ log. E Fe
			Quartz			
	P.R. 82–4	−0.81	N.D.	0.275	0.0028	−1.46
Boulder Creek	61–2	N.D.	N.D.	0.055	0.0083	N.D.
	56–2	N.D.	N.D.	0.039	N.D.	N.D.
	Peg. 83–2	N.D.	N.D.	0.017	N.D.	N.D.
	P.R. 113–4	−0.43	0.044	0.520	0.0310	N.D.
Overland Mountain	109–2	N.D.	N.D.	0.040	0.0079	N.D.
	Peg. 112–2	N.D.	N.D.	0.035	N.D.	N.D.
	117–3	−0.14	0.061	0.770	0.0500	0.16
	P.R. 100–3	−0.26	N.D.	0.510	0.0310	−0.01
Silver Plume	102–3	−0.19	0.031	0.700	0.0365	0.01
	Peg. 90–2	0.70	0.049	0.500	N.D.	N.D.
	89–2	N.D.	N.D.	N.D.	N.D.	N.D.

P.R.—Parent rock; Peg.—Pegmatite. N.D.—not detected.

TABLE 15.—*Quantitative analyses of muscovite (from pre-Cambrian granites)*

Granite	Sample number	% SrO	% BaO	% Sc_2O_3	Δ log. E La	% TiO_2	% V_2O_5	% Cr_2O_3	Δ log. E Mn
Boulder Creek	P.R. 81–6	0.100	0.100	0.0225	−0.64	1.80	0.105	0.0740	−1.34
	74–6	0.026	0.042	0.0120	−0.98	1.25	0.115	0.0420	−1.34
	Peg. 56–3	0.0042	0.0026	0.0195	N.D.	0.48	0.0083	0.0019	−1.38
	83–3	0.0068	0.0097	0.0740	N.D.	5.20*	0.018	0.0027	−1.26
Overland Mountain	P.R. 106–6	0.006	0.060	0.0220	N.D.	3.60*	0.080	0.0127	−1.15
	113–6	0.0135	0.0245	0.0280	N.D.	2.80*	0.060	0.023	−1.24
	Peg. 112–4	0.009	0.040	0.0150	N.D.	2.80*	0.095	0.0215	−1.35
Silver Plume	P.R. 117–5	0.007	0.012	0.0045	−0.60	1.10	0.041	0.0138	−1.76
	124–5	0.026	0.038	0.0100	−0.32	1.20	0.090	0.0225	−1.65
	102–5	0.013	0.028	0.0105	−0.73	1.20	0.056	0.0230	−1.79
	Peg. 90–3	0.0085	0.012	0.010	N.D.	0.67	0.0255	0.0021	−1.60
	89–3	0.0037	0.0052	0.0037	N.D.	0.92	0.020	0.0030	−1.60

P.R.—Parent rock; Peg.—Pegmatite. N.D.—not detected.

* Too high (read from standard curve extrapolated too far)

TABLE 16.—*Quantitative analyses of biotites*

Rock		Sample number	% SrO	% BaO	% Sc_2O_3	Δ log. E La	% TiO_2*	% V_2O_5	% Cr_2O_3	Δ log. E Mn	Δ log. E Co	% NiO	% ZnO
						Pre-Cambrian granites							
Boulder Creek	P.R.	58–5	0.135	0.022	0.036	N.D.	2.45	0.180	0.084	−0.38	−0.82	0.017	0.10
		82–5	0.054	0.017	0.0215	−0.51	2.75	0.150	0.160	−0.30	−0.60	0.033	0.12
		81–5	0.011	0.060	0.0165	−0.91	3.50	0.155	0.105	−0.70	−0.75	0.0155	N.D.
		54–5	0.0135	0.080	0.017	−0.63	7.40	0.120	0.160	−0.36	−0.58	0.019	0.24
		74–5	0.026	0.056	0.014	−0.67	5.60	0.120	0.120	−0.31	−0.57	0.027	0.185
	Peg.	59–3	0.0032	0.0115 0.052	0.0475	N.D.	5.00	0.107	0.006	−0.31	−0.74	0.017	0.008
Overland Mtn.	P.R.	106–5	0.011	0.052	0.018	−0.70	4.00	0.080	0.0127	−0.42	−0.85	0.0165	0.174
		113–5	0.0048	0.015	0.0105	−0.27	2.60	0.056	0.0245	−0.56	−0.80	0.015	0.150
	Peg.	112–3	0.0056	0.017	0.0054	N.D.	2.10 2.10	0.0465	0.0138	−0.66	−1.01	0.0023	0.105
Silver Plume	P.R.	117–4	0.0092	0.0083	0.0082	−0.93	2.10	0.062	0.0205	−0.66	−1.22	0.0026	0.063
		124–4	0.0182	0.015	0.0048	−0.30	4.10	0.078	0.035	−0.50	−0.88	0.0030	0.154
		100–4	0.0068	0.0095	0.0118	−0.35	4.80	0.078	0.038	−0.70	−1.04	0.0036	0.130
		102–4	0.0037	0.0074	0.0075	−0.52	3.0	0.0545	0.025	−0.92	−1.16	0.0026	N.D.
						Tertiary							
Granodiorite		35–5	0.058	0.050	0.003	−0.90	9.6	0.10	0.0085	−0.27	−0.57	0.012	0.21

P.R.—Parent rock; Peg.—Pegmatite. N.D.—not detected.

* Relative only (too high because read from standard curve extrapolated too far)

richest in BaO, followed in order by plagioclases, groundmasses, and micas. Al_2O_3 is frequently an abundant minor constituent in quartz. Some undoubtedly will find the percentages difficult to believe. Gravimetric analyses admittedly should be made. However, the author de-

TABLE 17.—*Minor constituents and minerals in which they occur (arranged in order of decreasing abundance for each constituent), based on quantitative analyses*

Mg	Ca	SrO	BaO
Groundmass Quartz	Micas Groundmass K-feldspars Quartz	Feldspars Groundmass Biotite Muscovite	K-feldspars Plagioclase Groundmass Biotite Muscovite Quartz
Al_2O_3	Sc_2O_3	La	TiO_2
Very high in all silicates	Muscovite Biotite Groundmass	Biotite Muscovite	Biotite Muscovite Groundmass Feldspar Quartz
V_2O_3	Cr_2O_3	MnO	Fe
Biotite Muscovite Groundmass Feldspars	Biotite Muscovite Groundmass Feldspars	Biotite Muscovite Groundmass	Groundmass K-feldspars
Co	NiO	ZnO	
Biotite	Biotite	Biotite	

fends the figures on the strength of the following facts: (1) The Al_2O_3 standard concentration curve for a quartz base was quite straight, showing that no error was made in preparing the standard mixtures, (2) some pegmatite samples (where picking was easy and possibilities for inclusion of feldspar grains were at a minimum) show appreciable Al_2O_3, (3) great care was used in picking parent rock samples, (4) the ability of Al to substitute for Si in the SiO_4 tetrahedra of silicate structures is well recognized by writers on crystal chemistry (Evans, Bragg, Wyckoff, Hassell, and Stillwell), and (5) all spectroscopic analyses of quartz (where Al was looked for), except one, with which the author is familiar, report the presence of Al. (Kennard, 1935; Doelter, 1914).

The micas contain most Sc_2O_3, the groundmasses less, the feldspars and quartzes none. La occurs only in the biotites and some muscovites. All biotites and most muscovites contain more than 1 per cent of TiO_2, the Ti

probably as a substitute for Mg and ferrous Fe. Figures for TiO_2 that exceed 1.2 per cent should be used only in a relative sense, as the spectrographic method is not very accurate above 1 per cent. This precaution applies only to the micas. V_2O_5, less abundant than TiO_2, has

TABLE 18.—*Minerals and their minor constituents (listed in order of decreasing abundance for each mineral), based on quantitative work*

K-feldspar	Plagioclase	Groundmass	Muscovite
BaO	SrO	TiO_2	TiO_2
SrO	BaO	SrO	V_2O_5
TiO_2	MnO & TiO_2	BaO & MnO	
MnO	V_2O_5	V_2O_5	Cr_2O_3
Cr_2O_3 & V_2O_5	Cr_2O_3	Sc_2O_3	Sc_2O_3
			BaO & SrO

Biotite	Quartz
TiO_2	Al_2O_3
ZnO & V_2O_5	BaO and TiO_2
SrO & BaO	
Sc_2O_3 & NiO	

(Mg, Ca, Fe, and Na not included as minor constituents.)

a similar distribution and is highest in the micas, especially biotite. There is an unmistakable tendency for V_2O_5 and Cr_2O_3 to occur in minerals high in ferric Fe. Cr_2O_3 is distributed in exactly the same manner as V_2O_5, although not so abundantly. Mn is highest in the micas (Δ log. E values are given in the mica analyses because Mn was too abundant in the mica base used to prepare the standard mixtures, and definite percentages could not be determined), where it substitutes for ferrous Fe. Biotite contains much more Mn than muscovite. Co, NiO, and ZnO occur only in the biotites, as substitutes for ferrous iron. Fe, Co, Mg, and Ca are discussed as elements because they were not included as oxides in the mixtures used to prepare the standard concentration curves. Tables 17 and 18 briefly summarize the general relations shown by the quantitative analyses.

POTASH FELDSPARS

Table 12 gives quantitative spectrographic analyses of potash feldspars, all from Colorado rocks. Table 19 gives the potash feldspar mineral names, based on refractive indices. The writer attempts in the following discussions to express the average content of every minor constituent for each mineral (including groundmass) in a certain rock as the mean of a number of analyses. The course was taken on the basis of the following facts: (1) The only representative chemical analysis

of a mineral or rock is the mean of all available analyses; (2) the method used was the simplest means of discussing lucidly the enormous volume of data at hand; and (3) the individual analyses in most cases do not differ greatly from the averages. Several large variations in the data (causes already discussed) may of course be found, but these should not be used to discount the major arguments set forth.

The distribution of the minor elements (as averages) in the parent rock potash feldspars, in order of decreasing abundance, is as follows:

For Mg:

Parent rock		Average Δ log. E
Granodiorite	(T)*	0.69
Boulder Creek granite	(PC)**	0.61
Silver Plume granite	(PC)	0.50
Granite-quartz monzonite	(T)	0.46
Overland Mtn. granite	(PC)	0.26

For Ca:

Parent rock		Average Δ log. E
Granodiorite	(T)	0.26
Boulder Creek granite	(PC)	0.22
Overland Mtn. granite	(PC)	0.15
Silver Plume granite	(PC)	0.09
Granite-quartz monzonite	(T)	0.03

* (T) indicates Tertiary age.
**(PC) indicates pre-Cambrian age.

For SrO:

Parent rock		Av. % of SrO
Granite-quartz monzonite	(T)	0.233
Granodiorite	(T)	0.139
Overland Mtn. granite	(PC)	0.130
Boulder Creek granite	(PC)	0.087
Silver Plume granite	(PC)	0.081

For BaO:

Parent rock		Av. % of BaO
Overland Mtn. granite	(PC)	0.164
Boulder Creek granite	(PC)	0.141
Granite-quartz monzonite	(T)	0.130
Silver Plume granite	(PC)	0.095
Granodiorite	(T)	0.090

For TiO_2

Parent rock		Av. % of TiO_2
Granodiorite	(T)	0.028
Granite-quartz monzonite	(T)	0.0267
Overland Mtn. granite	(PC)	0.026
Silver Plume granite	(PC)	0.0195
Boulder Creek granite	(PC)	0.0124

For MnO:

Parent rock		Av. % of MnO
Boulder Creek granite	(PC)	0.0132
Granodiorite	(T)	0.0068
Granite-quartz monzonite	(T)	0.0064
Silver Plume granite	(PC)	0.0064
Overland Mtn. granite	(PC)	0.0053

For Fe:

Parent rock		Average Δ log. E
Granodiorite	(T)	0.20
Granite-quartz monzonite	(T)	0.10
Boulder Creek granite	(PC)	0.05
Silver Plume granite	(PC)	−0.07
Overland Mtn. granite	(PC)	−0.34

The special purity (except for BaO and SrO) of the Overland Mountain potash feldspars can be attributed to the fact that the rock is pegmatitic. It has been shown that pegmatite minerals tend to be purer than normal. The Tertiary potash feldspars are relatively richer than those from the pre-Cambrian rocks in SrO, TiO_2, MnO, and Fe. The Tertiary samples are lowest in BaO. Potash feldspars from the most basic rocks, the

Boulder Creek granite and the Tertiary granodiorite, contain most Mg and Ca.

SrO:BaO ratios for the potash feldspars (based on averages) are as follows:

Parent rock		Parent rock SrO:BaO ratio	Pegmatite SrO:BaO ratio
Granite-quartz monzonite	(T)	1.79	
Granodiorite	(T)	1.54	
Silver Plume granite	(PC)	0.85	1.65
Overland Mtn. granite	(PC)	0.793	1.44
Boulder Creek granite	(PC)	0.617	1.84

(T) indicates Tertiary age. (PC) indicates pre-Cambrian age.

SrO is more abundant than BaO in the Tertiary potash feldspars but less abundant in the pre-Cambrian samples. The pegmatite ratios are far greater than those for the parent rock. Individual exceptions exist in Table 12, but averages are used here. This means that SrO increases greatly relative to BaO in pegmatite samples. The author explains this as follows: Parent-rock potash feldspars formed at higher temperatures than those in pegmatites, and therefore in the former the structures were disordered and more tolerant to substitutional ions. This accounts for decreases in both BaO and SrO in the pegmatite mineral. At higher temperatures the feldspar structure is tolerant to both Sr and Ba as substitutes for K (ionic radii: Ba 1.43 Å; K 1:33 Å; Sr 1.27 Å), but at the lower pegmatite temperatures the tolerance for the smaller Sr ion must be greater than for the larger Ba ion, a fact that explains the lower ratios for the parent-rock potash feldspars. Since the ratios are lower for the lower-temperature feldspar, and since the order of increasing ratios is also the order of decreasing geologic age for the Jamestown district, the potash feldspars of the oldest rock in the district may have crystallized at the highest temperature, and those of succeedingly younger rocks at successively lower temperatures. This is merely a suggestion, but it agrees well with the field facts.

Parent rock-pegmatite relations for the various minor elements in the potash feldspars (averages) are as follows:

	Boulder Creek granite		Overland Mtn. granite		Silver Plume granite	
	P.R.*	Peg.**	P.R.	Peg.	P.R.	Peg
Av. Δ log. E Fe	0.05	−0.51	−0.34	−0.45	−0.07	−0.66
Av. Δ log. E Mg	0.61	−0.02	0.26	−0.24	0.50	0.24
Av. Δ log. E Ca	0.22	−0.14	0.15	0.07	0.09	−0.09
Av. % of SrO	0.087	0.044	0.13	0.128	0.081	0.061
Av. % of BaO	0.141	0.0239	0.164	0.089	0.095	0.037
Av. % of MnO	0.0132	0.0091	0.0053	0.0029	0.0064	0.0050
Av. % of TiO_2	0.0124	N.D.	0.027	0.0035	0.0195	N.D.

* Parent rock.
** Pegmatite.

The pegmatite potash feldspars average less of each minor constituent than the average for the parent-rock feldspar. These facts substantiate the conclusions drawn from qualitative results and have been explained on the basis of crystal structure.

TABLE 19.—*Compositions of feldspars from refractive indices*

Sample	Rock	Composition
	Potash feldspars	
124–2	Silver Plume granite	microcline
102–2	Silver Plume granite	orthoclase
90–1	Silver Plume granite (pegmatite)	microcline
89–1	Silver Plume granite (pegmatite)	microcline
74–3	Boulder Creek granite	microcline
82–3	Boulder Creek granite	microcline
58–3	Boulder Creek granite	microcline
61–1	Boulder Creek granite (pegmatite)	microcline
83–1	Boulder Creek granite (pegmatite)	microcline
106–3	Overland Mountain granite	microcline
113–3	Overland Mountain granite	microcline
109–1	Overland Mountain granite (pegmatite)	microcline
40–3	Tertiary granodiorite	orthoclase
33–3	Tertiary granodiorite	orthoclase
116–3	Granite-quartz monzonite	orthoclase
9–2	Granite-quartz monzonite	orthoclase
	Plagioclases	
74–2	Boulder Creek granite	An_{42}
81–2	Boulder Creek granite	An_{42}
50–2	Tertiary granodiorite	An_{45}
87–2	Tertiary granodiorite	An_{35}
106–2	Overland Mountain granite	An_{40}
113–2	Overland Mountain granite	An_{28}

PLAGIOCLASES AND GROUNDMASSES

Table 13 lists the plagioclase and groundmass quantitative analyses. Table 19 gives the compositions of the plagioclases as determined by refractive indices. Figure 1 locates the samples.

The distribution of the minor constituents in the plagioclases, again using averages, is as follows:

	Boulder Creek granite	Overland Mtn. granite	Tertiary granodiorite
Av. % of SrO	0.066	0.045	0.120
Av. % of BaO	0.034	0.060	0.011
Av. % of TiO_2	0.008	0.008	0.023
Av. % of MnO	0.010	0.004	0.013
SrO:BaO ratio	1.94	0.75	10.9
"Purity"	Intermediate	Purest	Most impure

The Tertiary plagioclase is exceptionally high in SrO, as was the Tertiary potash feldspar. BaO is most abundant where SrO is lowest, and vice versa. The Tertiary plagioclase had by far the highest SrO:BaO ratio. Tertiary potash feldspars also have larger ratios than the pre-Cambrian. Plagioclase ratios are far greater than those for potash feldspars, as would be expected from the especially strong affinity of Sr for Ca.

All minor constituents except BaO are most abundant in the granodiorite, next most abundant in the Boulder Creek granite, and least abundant in the Overland Mountain granite. This holds not only for the constituents listed above, but also for V_2O_5 and Fe. On this basis the "index of purity" for the plagioclases was included. The purity of the Overland Mountain granite samples is very probably due to its pegmatitic nature.

The content of MnO, Cr_2O_3, V_2O_5, and TiO_2 is usually large in the groundmasses of the porphyritic rocks, and Sc_2O_3 is usually present.

MUSCOVITES

Table 15 gives quantitative analyses of muscovites, all from the pre-Cambrian rocks. The average percentages of the various minor constituents in the muscovites from the parent rocks and pegmatites are as follows:

	Boulder Creek granite		Overland Mtn. granite		Silver Plume granite	
	P.R.*	Peg.**	P.R.	Peg.	P.R.	Peg.
Av. % of SrO	0.063	0.0055	0.0098	0.009	0.015	0.0066
Av. % of BaO	0.071	0.0062	0.042	0.040	0.026	0.0086
Av. % of Sc_2O_3	0.017	0.046	0.025	0.015	0.008	0.0069
Av. % of TiO_2	1.52	erratic	3.20	2.80	1.2	0.80
Av. % of V_2O_5	0.11	0.013	0.07	0.095	0.062	0.023
Av. % of Cr_2O_3	0.058	0.0023	0.02	0.0215	0.0178	0.0026
SrO:BaO ratio	0.89	0.89	0.23	0.23	0.58	0.77

* Parent Rock.
** Pegmatite.

La is present in the Boulder Creek and Silver Plume parent rock samples and absent in the pegmatite samples. Mn is most abundant in the Boulder Creek and Overland Mountain muscovites. Little or no decrease occurs for Mn in pegmatite as compared to parent-rock muscovites.

The Boulder Creek and Silver Plume pegmatite samples bear less SrO, BaO, TiO_2, V_2O_5, and Cr_2O_3, on the average, than those from parent rocks. MnO and Sc_2O_3 seem to be distributed at random. The data for the Overland Mountain granite are too meager to permit conclusions, but because of its pegmatitic nature no great differences would be expected in the parent-rock muscovites as compared to those from pegmatites. The differences between the pegmatite and parent-rock muscovites are due to the same conditions described for the potash feldspars.

All SrO:BaO ratios are less than unity. (In all cases BaO averages more than SrO.) Pegmatite ratios are similar to those for the parent rocks. This might be explained on the basis that for muscovites there is no pronounced inversion from a disordered to an ordered structure with lowering of temperature. This is purely hypothetical. A pronounced inversion probably does take place in the case of potash feldspars.

The Boulder Creek muscovites are highest in average content of SrO, BaO, V_2O_5, and Cr_2O_3. The Silver Plume samples contain the least of all minor constituents investigated.

BIOTITES

The biotite analyses appear in Table 16. The average percentages of the various minor constituents in the biotite samples are as follows:

	Tertiary granodiorite	Boulder Creek granite	Silver Plume granite	Overland Mtn. granite
Av. % of SrO	0.058	0.048	0.0095	0.0079
Av. % of BaO	0.050	0.047	0.010	0.034
Av. % of Sc_2O_3	0.0030	0.021	0.0081	0.019
Av. % of TiO_2*	9.60	4.30	3.50	3.30
Av. % of V_2O_5	0.10	0.145	0.068	0.068
Av. % of Cr_2O_3	0.0085	0.126	0.030	0.019
Av. % of NiO	0.012	0.022	0.003	0.016
Av. % of ZnO	0.21	0.16	0.116	0.163
Av. Δ log. E La	−0.90	−0.68	−0.53	−0.48
Av. Δ log. E Mn	−0.27	−0.41	−0.69	−0.49
Av. Δ log. E Co	−0.57	−0.66	−1.10	−0.83

		P.R.	Peg.	P.R.	Peg.	P.R.	Peg.
SrO:BaO ratio	1.15	1.0	0.26	0.95	....	2.3	0.33

* Figures relative only. Percentages much too high.

The two pegmatite samples analyzed contain less of all constituents (except Sc_2O_3, TiO_2, and Mn) than the averages for their parent rocks. The explanation is the same as that given previously for potash feldspars. The Silver Plume biotites bear less ZnO, NiO, Co, Mn, V_2O_5, and BaO than the biotites from the other rocks and are low in SrO, as was also

true for the muscovites from the same rock. Since the Silver Plume micas are the purest of the group, an interesting question arises. Can one determine by the spectrographic method which rocks are normal intrusives and which are migmatites? The writer believes that the Silver Plume is the only granite of the pre-Cambrian series not related to hydrothermal origin. The Overland Mountain mass has been described as peculiarly pegmatitic and the Boulder Creek granite varies from true granite to granitized schist. Should the minerals from an intrusive rock be purer from the standpoint of minor constituents than the minerals in migmatites? Is the reverse true?

QUARTZ

Table 14 lists the quartz analyses. Quartz contains relatively few minor constituents. The average percentages of the several constituents found in the quartz samples are as follows:

	Boulder Creek granite		Overland Mtn. granite		Silver Plume granite	
	P.R.*	Peg.**	P.R.	Peg.	P.R.	Peg.
Av. Δ log. E Mg	−0.81	N.D.***	−0.43	N.D.	−0.20	Minute
Av. % Al_2O_3	0.275	0.027	0.52	0.037	0.66	0.5 and absent
Av. % TiO_2	?????	?????	0.031	0.0079	0.0392	N.D.

* Parent rock.
** Pegmatite.
*** Not detected.

The Mg and Al_2O_3 content of the quartz increases as the age of the parent rock decreases. Pegmatite samples are purer than those from the parent rocks. The TiO_2 content is interesting in that microscopic rutile needles often seen in quartz might be due to separation from a single phase with lowering of temperature.

SUMMARY AND CONCLUSIONS

Qualitative and quantitative spectrographic analyses of a large number of igneous rocks and minerals from the Jamestown district, Colorado, showed that the minor elements were systematically distributed. Several Massachusetts rocks analyzed for comparison brought out the importance of the province relationship.

Conclusions can be stated as follows:

(1) From qualitative analyses of rocks, a scheme may be set up whereby any rock in the district can be identified on the basis of its minor elements. Quantitative analyses of minerals permit a further, more reliable, division and identification scheme. Correlation of rocks within

a district is therefore possible by use of the spectrographic method. Correlation of rocks between districts may also be possible.

(2) Each mineral contains characteristic minor constituents, which are definitely related to the major constituents present. In most cases the ions of minor elements occur as substitutions (solid solution) for ions of the appropriate-sized major elements.

(3) The SrO:BaO ratios are highest in potash feldspars from the youngest rock and lowest for those from the oldest rock. This was not true for all the minerals but may nevertheless be significant. Plagioclases have the highest SrO:BaO ratios, and muscovites the lowest. Pegmatite potash feldspar ratios are higher than those for parent rock potash feldspars. This is explained by crystal structure-temperature relations.

(4) A most important principle resulting from this work is: Each pegmatite mineral averages less of almost every minor constituent than the average for the same mineral from the parent rock. A rational explanation can be given on the basis of crystal structure.

(5) Other factors aside from structure-temperature-ionic radius relations that affect the quantity of minor elements found in the minerals are: (1) the relative abundance of these constituents in the original magma, granitizing solutions, or rock being granitized, and (2) the pressure conditions.

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